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VOLUME 60 October 1949 NUMBER 10

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PAPERS IN PRESS FOR FORTHCOMING ISSUES

RECONNAISSANCE GEOLOGY AND VERTEBRATE PALEONTOLOGY OF EL SALVADOR, CENTRAL AMERICA. By R. A. Stirton and W. K. Gealey

VALLEY-WIDENING PROCESSES IN THE COLORADO ROCKIES. By B. A. Tator

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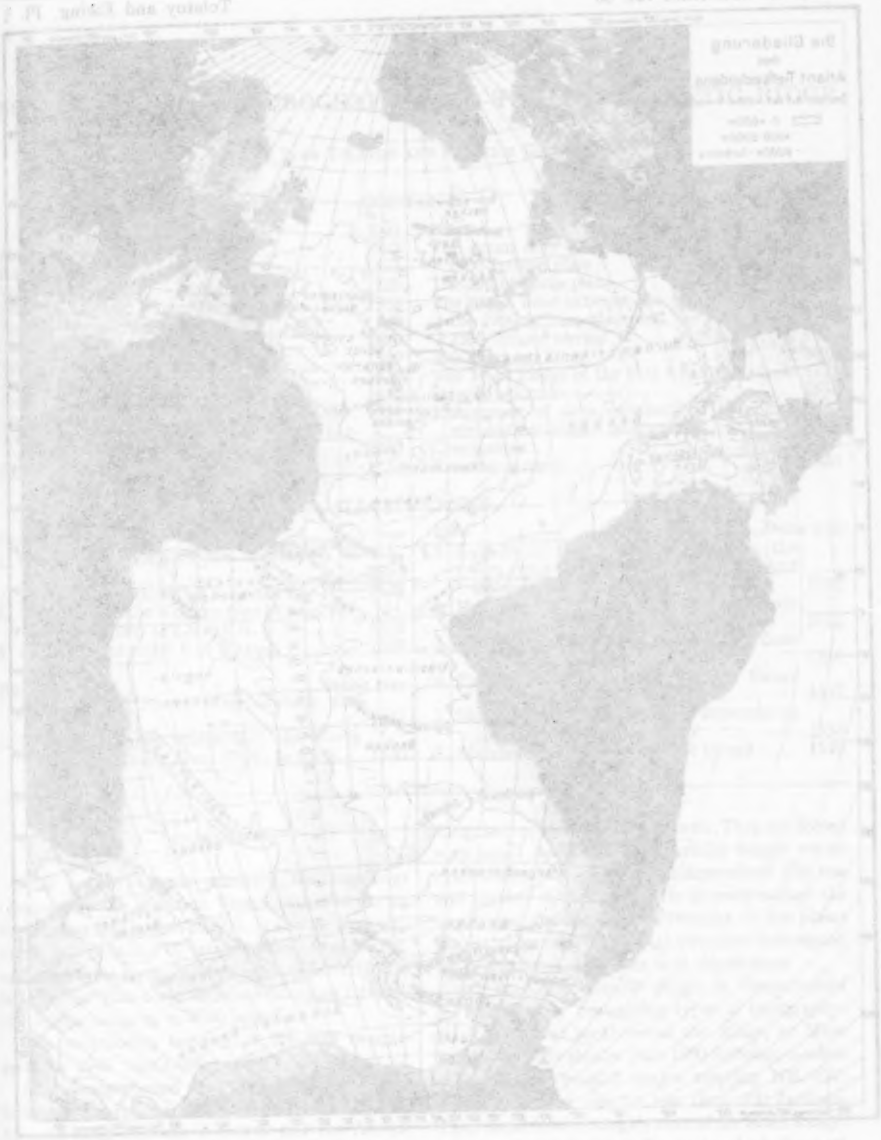
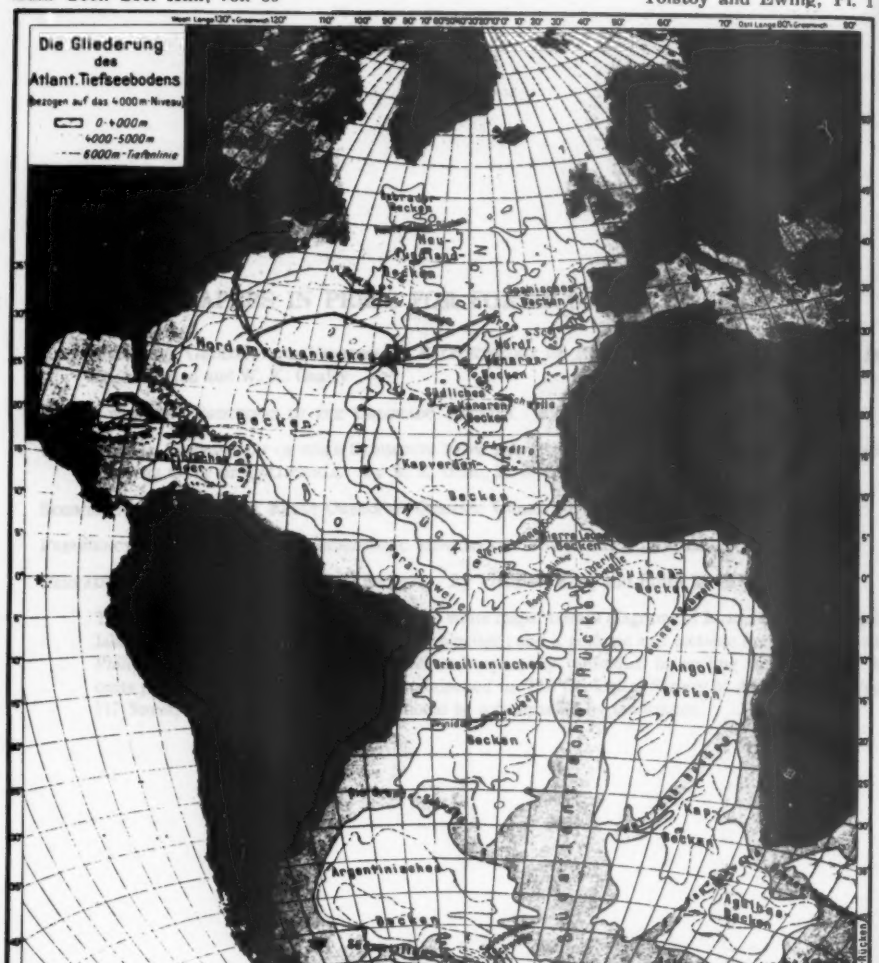
SEISMIC EVIDENCE FOR THE FAULT ORIGIN OF OCEANIC DEEPS. By Hugo Benioff

FREQUENCY AND PERIODICITY OF OROGENETIC MOVEMENTS. By L. M. R. Rutten

TERRAIN DIAGRAMS OF THE PHILIPPINE ISLANDS. By Philip B. King and Edith M. McKee

This paper is accompanied by five black and white maps: terrain diagrams of Luzon and the Visayan Islands, Mindanao, Palawan, and Sulu Archipelago; and a geologic and tectonic sketch map of the Philippine Islands. These may be purchased separately (unfolded, in mailing tubes). Price 50 cents per set of five. Orders should be addressed to The Geological Society of America, 419 West 117 Street, New York 27, N. Y., and should be accompanied by remittance.





THE ATLANTIC OCEAN, AFTER MAURER AND STOCKS

The path of the ATLANTIS during the summer of 1947 (Cruise 150) is shown by the black lines

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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NORTH ATLANTIC HYDROGRAPHY AND THE MID-ATLANTIC RIDGE

By Ivan Tolstoy and Maurice Ewing

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Abstract

This paper presents primarily the soundings obtained by the Research Vessel Atlantis during the summer field season of 1947. It also utilizes some of the data obtained by later expeditions of the Atlantis and by several other vessels. The Atlantis is equipped with a continuously recording fathometer with a range up to 4000 fathoms.

The outstanding features of the area investigated in 1947 (Atlantis cruise #150) are:

(1) A conspicuously flat plain at a depth of 2900 fathoms occupies the deeper sections of the North Atlantic basin between Bermuda and the Azores. More recent data indicate that this horizontal stretch of sea floor extends at least as far south as Lat. 29°N. and possibly as far north as Lat. 40°N., between Long. 50°W. and 56°W., and also that at the same depth a similar plain is found east of the Ridge in the Northern Canary basin. They are dotted with small sea mounts of variable height whose exact shapes have not yet been determined. The size and number of these appear to increase toward the south, and the southern extremities of the plains consist of smooth horizontal stretches interrupted by pronounced elevations and depressions.

(2) The Mid-Atlantic Ridge is characterized by two strongly contrasting types of topography: (a) The central backbone of the Ridge, or Main Range, which is shoaler than 1600 fathoms, consists of a series of parallel ranges trending NE.–SW. Several of these rise to less than 800 fathoms. Their trend follows roughly that of the Main Range. The portion of the Main Range to which the most attention was devoted is between Lat. 30° and 34°N. In this area its width is of the order of 150 miles, and it is of the order of several hundred miles in the area of the Azores.

(b) The flanks, between the 1600- and 2500-fathom isobaths, consist of a succession of smooth shelves, each from 1 to 50 nautical miles or more in width. This zone is 200–300 miles in width.

(c) The region between the limit of the 2900-fathom plain and the foot of the first of the series of shelves characterizing the flanks of the Ridge, at 2500 fathoms, in some localities stands out as a distinct physiographic province.

(3) The following local features are noteworthy:

(a) Close to 31°N.Lat. a deep east-west trench extends from about 41° to 43°W.Long. and cuts deep into the Main Range. Its deepest point is at 2800 fathoms. Crushed and metamorphosed ultrabasics were brought to the surface by dredging its flanks.

(b) At 30°15′W., 34°N., a flat-topped sea mount rises to 180 fathoms, about 100 miles southeast of the Main Range. The dredge brought up calcareous discs of probably Cenozoic age off its top. They were about 15 cm in diameter and about 4 cm thick.

(c) Another sea mount was found at 33°43′N., 62°30′W., 150 miles NE. of Bermuda, shoaling to 780 fathoms, having a NW.-SE. elongation, an unknown total length, and a width of 4 miles across its flat top.

(d) Another flat suboceanic plain occupies the bottom of the basin separating Bermuda from the Northeastern United States, at a depth of 2650 fathoms.

Introduction

The authors' intention is to bring as rapidly as possible to the public's attention the existence of certain very important features of ocean-bottom topography. It is hoped that they will attract the attention of investigators to certain crucial points which can to a large extent be elucidated by a more widespread use of recording fathometers with a range of 0 to 4000 fathoms.

The present paper is the first of a series which will describe the data obtained by the Atlantis cruises in the North Atlantic since the summer of 1947. It was judged sufficient to illustrate it by means of maps and fathograms. Of course, the latter provide only somewhat distorted profiles of the sea floor. However, this distortion is always small enough so that all the striking features stand out very satisfactorily. Part II of this paper will give a series of corrected profile drawings.

Since the 1947 summer cruise (cruise #150) the Atlantis has made three cruises across the North Atlantic (cruises 151, 152, 153). Also, three almost uninterrupted fathograms of other ships crossing the Atlantic, as well as several shorter ones and numerous discrete sounding lines, were made available for examination, thanks to the kindness and co-operation of the officers of the U. S. Navy Hydrographic Office. Some of the main features of these data will be mentioned here in passing. Their detailed presentation is reserved for a later paper.

Acknowledgments

This work was made possible through the support of the National Geographic Society, The Woods Hole Oceanographic Institution, The Geological Society of America, and the Bureau of Ships, Navy Department. Most of the data were obtained by the Research Vessel Atlantis, belonging to the Woods Hole Oceanographic Institution. The authors are very grateful to Dr. Columbus O.'D. Iselin for facilitating in many ways the task of obtaining data from cruises other than those with which the authors were directly connected. The authors are also much indebted to Dr. J. B. Hersey for the loan of his data obtained on Atlantis cruises prior to cruise 150. Through the courtesy and co-operation of the U. S. Navy Hydrographic Office many additional data have been made available to the authors. It is expected that in co-operation with this institution the authors will soon obtain a much more complete and satisfactory picture of the North Atlantic Ocean bottom. Finally the co-operation and work of Adrian Lane, Master of the Atlantis, and of his crew were greatly appreciated at all stages of the field work.

The Mid-Atlantic Ridge: Generalities

The Mid-Atlantic Ridge (or Swell, or Rise) has attracted much attention since the end of the nineteenth century. It is the most conspicuous feature of the Atlantic Ocean basin and is essentially a long narrow submarine mountain range which extends from about Lat. 65°N. (Iceland) to Lat. 50°S. It parallels the American, European, and African coasts (Pl. 1). At the equator this parallelism is strikingly illustrated by a pronounced E.-W. flexure.

Plate 1 shows the axis of the Ridge to be al-

most entirely above 2240 fathoms (4000 m), and bordered by deeper regions which are themselves split up into a series of basins, whose depths are of the order of 3000 fathoms, by ridges branching out from the main body of the Mid-Atlantic Ridge. These diverging mountain ridges give a herringbone-like aspect to the relief of the Atlantic floor, particularly in the southern hemisphere. The high parts of the Ridge shoal to less than 1000 fathoms in places, and the highest peaks form groups of islands such as the Azores and St. Paul's Rocks. North of Lat. 45°N. the Mid-Atlantic Ridge broadens and rises gradually to Iceland. South of Lat. 50°S. the Ridge bends eastward, parallel to the South African coast, and is believed by some authors (Wiseman and Seymour Sewell, 1937) to extend to the Carlsberg Ridge in the Indian Ocean.

Previous Hydrographic Work

The map shown in Plate 1 was published in the account of the Meteor oceanographic expedition of 1925–1927 (Maurer and Stocks, 1933), during which 14 echo-sounding profiles, with soundings spaced every 20 minutes of time, were run from coast to coast between Lat. 20°N. and 50°S. The Meteor (Dietrich, 1939) and Altair (Wüst, 1939) did more surveying in the North Atlantic in 1934–1939, on the ridge north of the Azores and on the Azores plateau itself. The U.S.S. Muir made two crossings of the Atlantic: from Philadelphia to Dakar and back in 1945, taking soundings every half hour.

All these soundings have contributed greatly to our knowledge of the bathymetry of the Atlantic Ocean basin and have in particular helped to emphasize the distribution of submarine mountain ranges with respect to the deeper basins. However, these surveys consisted of a series of discrete soundings. Until now no continuous fathograms of any portion of the Mid-Atlantic Ridge have been available. As will be shown, important features of ocean-bottom structure may escape notice if continuous fathograms are not available.

Gravity Measurements

Vening-Meinesz (1934) has shown that the Mid-Atlantic Ridge does not seem to be associated with any conspicuous gravity anomalies. However, the isostatic anomalies tend to be positive, particularly in the area of the Azores, where the regional isostatic anomalies are the smallest, showing that the compensation of the Azores plateau, like that of many other island blocks, is regional.

Seismicity

Though this is one of the principal seismic belts of our globe, its activity is in no way comparable with that of the Circum-Pacific Belt. All the shocks known to have originated in the area of the Mid-Atlantic Ridge are of the shallow type and are mostly of the c and d classes (Gutenberg and Richter, 1949). There seem to be three regions on the Ridge in which seismic activity is concentrated: the eastern extremity of the equatorial flexure, in the vicinity of St. Paul's Rocks; the western extremity of the flexure, around 10°N. 40°W.; and in the vicinity of the Azores. The portion of the Ridge between Lat. 25° and 35°N. also seems to be somewhat more active than the rest of the Ridge, though this may be due to the favorable location of these epicenters with respect to the existing seismological stations. The first three localities mentioned have produced much larger shocks than have been noted elsewhere on the Ridge—class b and even a (Azores, Nov. 25, 1941). The Mid-Atlantic Ridge is thus the locus of a certain amount of unreleased stress, which seems to be mostly concentrated at the two ends of the equatorial flexure and also in the vicinity of the Azores plateau.

Theories on the Origin of the Ridge

There has been much speculation as to the origin of the Mid-Atlantic Ridge. Haug (1900) believed it to be the median anticline of what he called the Mid-Atlantic "geosyncline".

Taylor (1910) considered it as a horst. Wegener (1924) believed it to be the bottom of the rift that was opened up prior to the separation and drifting apart of the bordering continents. Molengraaff (1928) was of essentially the same opinion. H. S. Washington (1930) stresses the geological significance of the ultrabasics of St. Paul's Rocks. These rocks are peridotites that indicate crushing and meta-

morphism by pressure. He regards them as evidence of lateral pressure in the building of the Ridge. He also remarks that Prinz's theory of differential torsion of the two hemispheres might explain the prevalent trends of the Ridge, and in particular the equatorial flexure.

Bucher (1940) believes in the essential similarity of the pattern of continental and oceanic structures in the North Atlantic and surroundings areas. He says the pattern of the ocean floor seems to be "comparable to that of the basins and swells of the continental areas outside the orogenic belts, although the scale is both larger horizontally and vertically on the oceanic surfaces."

Umbgrove (1947) emphasizes the symmetry of the distribution of basins and ridges in the South Atlantic and in the African continent and suggests that the prevalent trends are here undoubtedly those of the two major linear patterns of the earth's crust (the N.–S., E.–W., and bisecting systems). Hans Cloos (1939) on the basis of an experiment on clay has shown that the type of topography characterizing the Azores plateau can be explained in terms of doming of a section of the earth's crust with formation of fractures parallel to the longest dimension of the domed section (NW.–SE.). Subsequent extrusion of magma through fissures would have formed the NW.–SE. ridges surveyed by the ALTAIR.

Agostinho (1936) also believes the major topographic features of the Azores plateau are due to the extrusion of volcanics through predominantly northwest-southeast fractures. In addition he thinks the origin of this big bulge in the Ridge is somehow related to the fact that it is situated at what he considers the intersection of two major submarine trends: the almost north-south trend of the Mid-Atlantic Ridge and the almost east-west trend formed by the succession of banks beginning with the Great Banks, through Milne Bank, the Azores, and the banks west of Gibraltar.

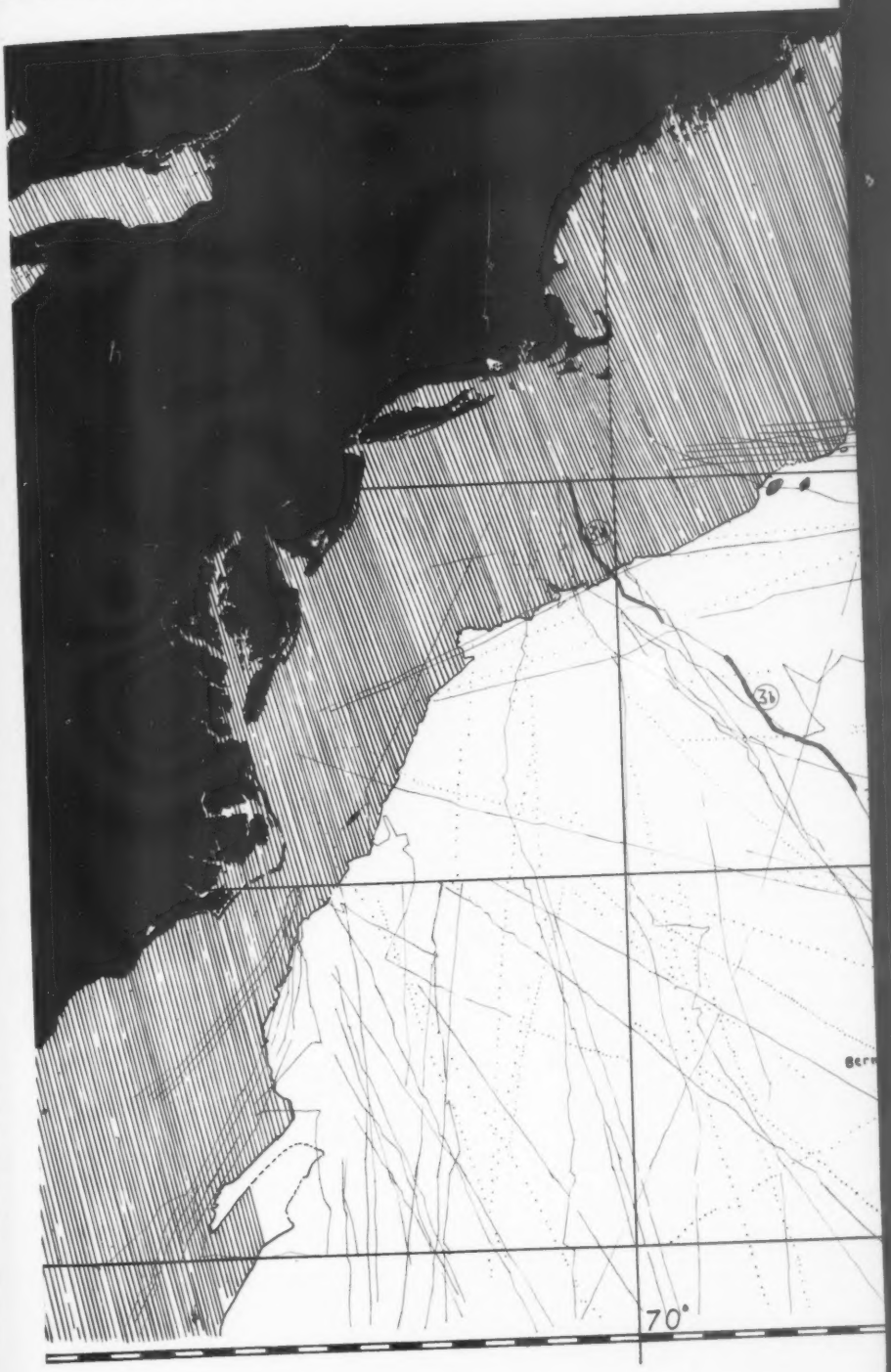
Thus the status of the problem of the origin of the Mid-Atlantic Ridge is somewhat confused. The present authors are not prepared to take any definite stand on this aspect of the problem. However, the possible implications of the results obtained so far are indicated, since our investigation has revealed a series of hitherto unsuspected features of the ocean floor, both on the Mid-Atlantic Ridge itself and in other parts of the North Atlantic.

ACCURACY OF DEPTHS AND POSITION

Contours of the ocean bottom were drawn where such a procedure was judged practicable —*i.e.*, when intersections between various parts of the ship's track occurred, or when the distance between two tracks was sufficiently small. No absolute criteria for the justification of such a procedure can be established, since the determining factors, such as the type of topography and accuracy of navigation, are subject to considerable variations from place to place and from day to day. Accuracy of the topographic maps varied and depended upon numerous factors, the relative importance of which cannot always be determined. The following causes of error should be noted, and kept in mind when considering the results here reproduced.

(1) The ship's position was obtained by astronomical methods and dead reckoning. An accuracy of ± 1 mile is a reasonable estimate for the ship's position as obtained in the evening and morning from star fixes, since especial care was taken. The noon fix should not be relied upon to better than ± 2 miles. The largest possible errors would occur for intermediate times.

The effect of currents and wind on a small ship, cruising at 8 knots, can be very great and had to be considered several times in the drafting of the hydrographic maps. Such corrections are made possible by the closing, at one or several points, of the network of ships' tracks. Accuracy is proportional to the number of ships' tracks available, particularly when associated with star fixes. Such corrections were made by mutual adjustment of the ship's supposed courses drawn on tracing paper, which process was continued until all intersections were made consistent with the soundings. In many cases the necessary adjustments were small, and in several cases unnecessary. Several times when some interesting feature was encountered, such as an unusually high ridge, box surveys were conducted. A zigzagging course was adopted, each change in course being determined from the readings on the fathogram, in order to try to enclose the feature within a network of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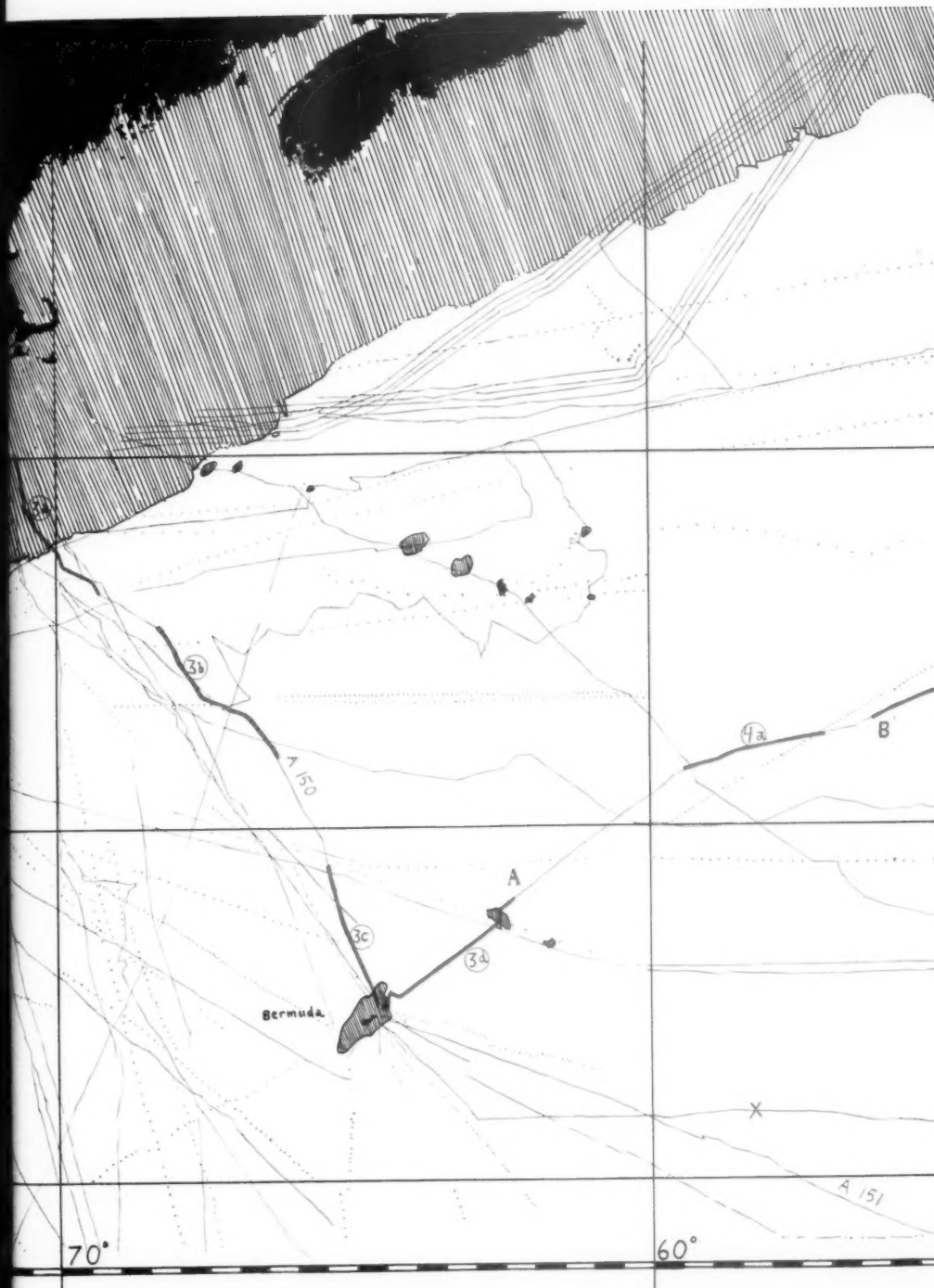
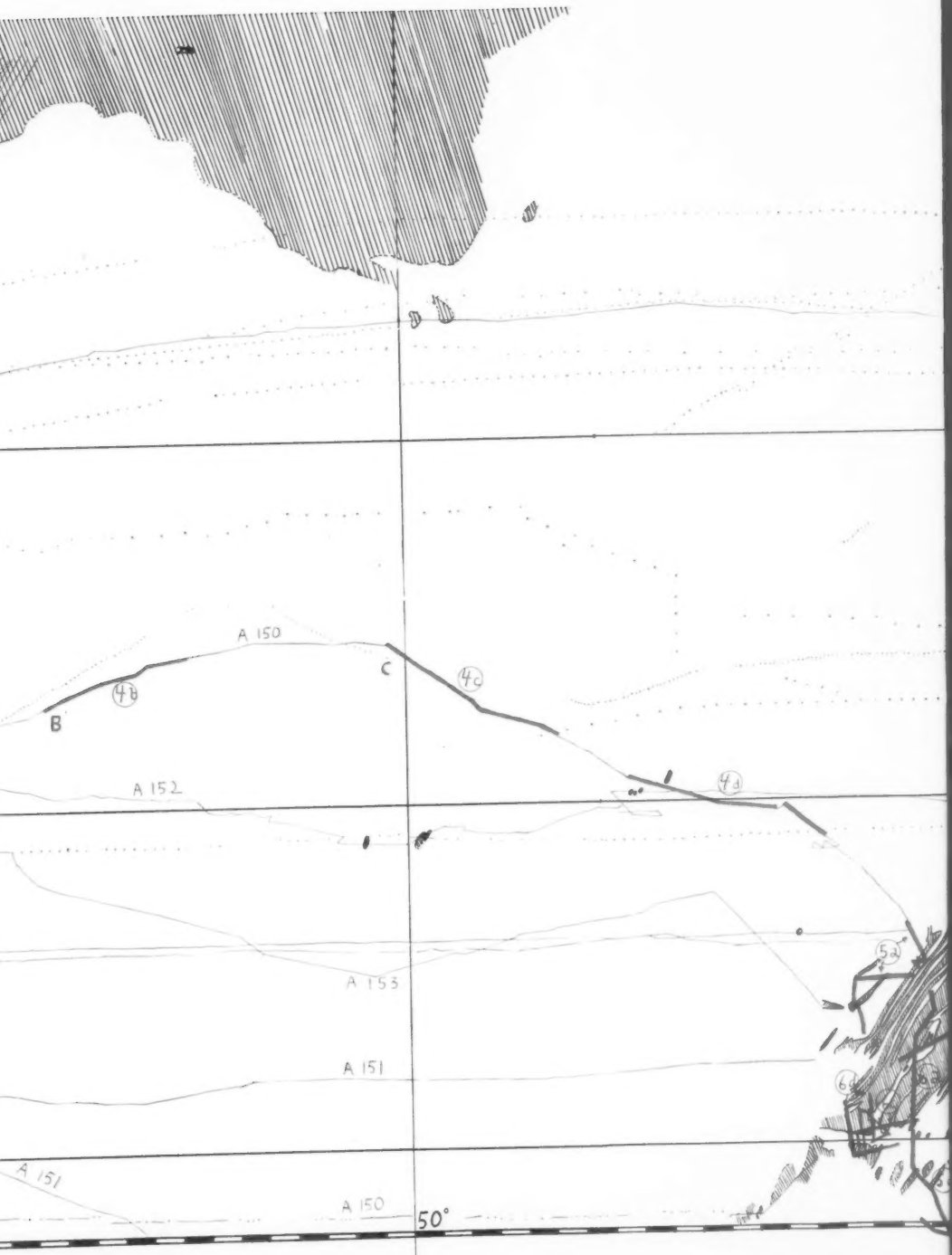
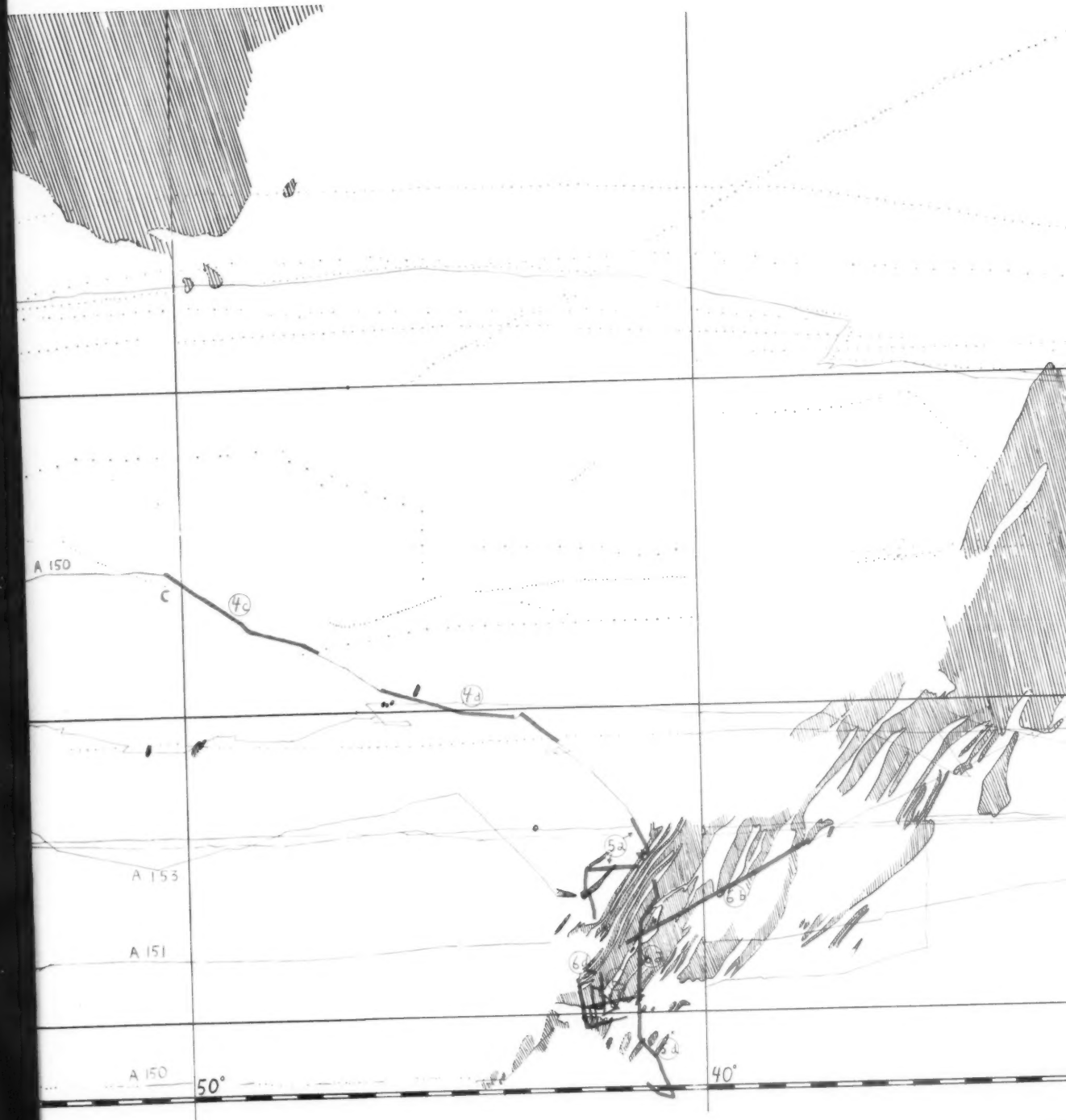
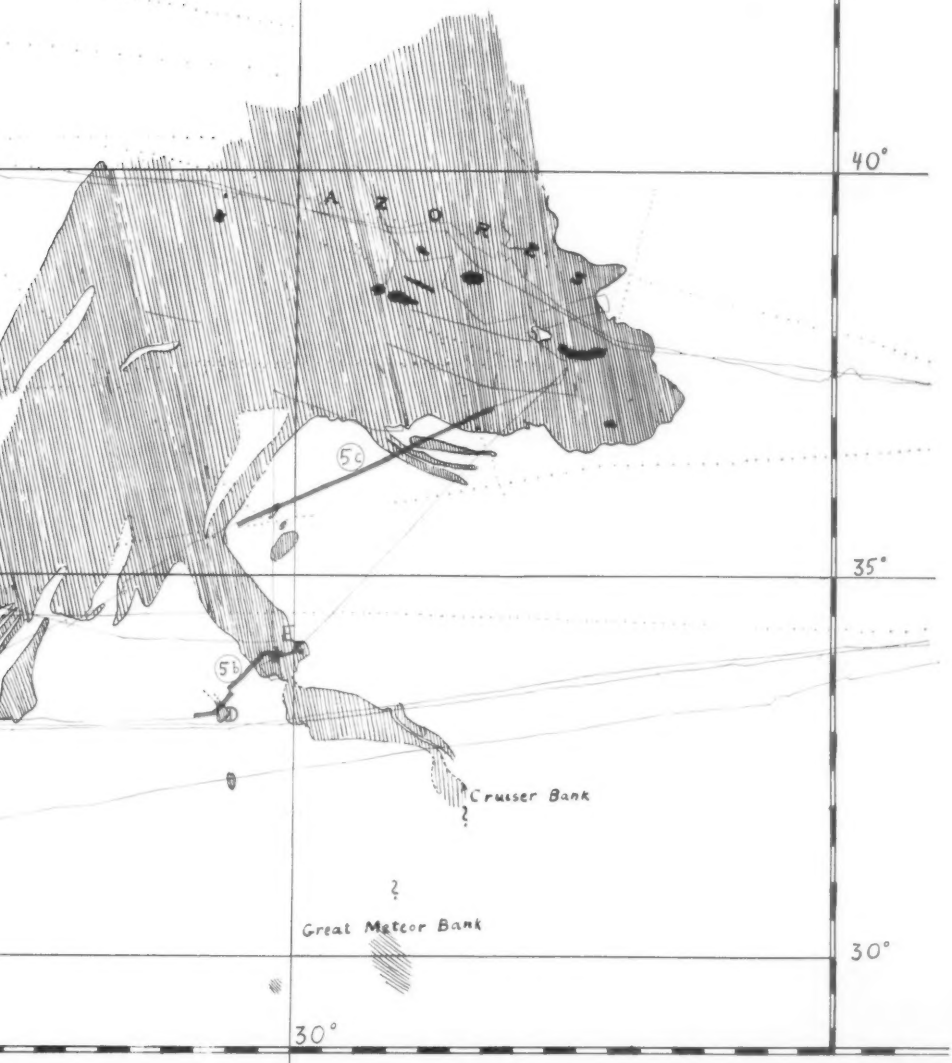

3a
3b
A 150
4a
B
A
3c
3d
Bermuda
X
A 151
70°
60°


A 150
B
C
4b
4c
4d
A 152
A 153
A 151
A 151
A 150
50°
5a
6a


A 150
C
4c
4a
5a
6b
6a
A 153
A 151
A 150
50°
40°

THE MID-ATLANTIC RIDGE AND THE NORTH AMERICAN BASIN, FROM 29° TO 45°N.

Areas shoaler than 1600 fathoms are shown by hachures. Red Lines indicate the sounding lines utilized so far. The continuous lines indicate either that fathograms are available, or that the soundings taken were sufficiently closely spaced to be equivalent for all practical purposes to a continuous record. The thick lines indicate the locations of the fathograms reproduced with the corresponding numbers and letters in Plates 3, 4, 5, and 6. The numbers A150, A151, A152, A153 indicate the corresponding ATLANTIS cruises across the Atlantic.



shortest possible tracks. Accuracy of positioning of the soundings increases with the number of intersecting courses and with more rugged topography. However, absolute geographical positioning will remain open to errors of a larger order of magnitude, except when there are one or several star fixes in the area. These are always fixed points in the transformations of the adjustment.

(2) The errors of sounding may be due to three causes:

(a) Departure from 60 cycles of the frequency of the generator supply to the synchronous motor running the stylus arm of the recording apparatus. The vertical divisions of the recording tape are calibrated to give exact depth readings for a veolocity of the stylus arm corresponding to 60 cycles. A departure from this figure of ± 1 cycle will give a reading which will be in error by $\pm\frac{1}{60}$ of the true depth, which at a depth of 2000 fathoms would mean an absolute error of 33 fathoms. Most of the recording is free from errors of ± 1 cycle or more—or, at least, these will usually be readily detectable on the fathogram. However, since there was no automatic frequency-control device, changes of the order of half a cycle may have been fairly common and may have been overlooked. Therefore, in oceanic depths the soundings should not be relied upon to better than ± 20 fathoms.

(b) The angle at the apex of the outgoing cone of sound, which leads sometimes to the so-called slope correction (Adams, 1942). This error is not sufficiently large to be taken into account in the plotting of the maps. Thus if a vertical wall were supposed to rise from the ocean floor to sufficient height from a depth of 2000 fathoms, and if we suppose the half angle at the apex of our cone of sound to be of the order of 25°, the error in the positioning of the foot of the slope would be of 12000 sin 25° = 5000 feet approximately. If this wall were 1000 fathoms high, it would therefore record as a slope of $\tan^{-1}\frac{6}{5}$ or 50° approximately. Now since 5000 feet is always an error of lesser magnitude than those involved in dead-reckoning navigation, and since the steepest continuous average slopes of any length that were recorded were of the order of 25°–30°, there is little reason for applying this correction to the drafting of our topographic maps. We must just keep in mind that the slopes shown on the fathograms are a little less steep than the true slopes, and that vertical cliffs such as might be expected to occur in connection with long continuous rises of many thousands of feet over a few miles will not show up as such on fathograms.

(c) Variations in the velocity stratification of the ocean. The errors due to this cause are not very large (of the order of 1 per cent of the total depth or less). We here follow the general policy of such centers of hydrographic study as the U. S. Hydrographic Office and the U. S. Coast and Geodetic Survey and disregard them.

General Description of the Areas Investigated

Plate 2 shows the various sounding lines now available. They include Atlantis cruises Nos. 143, 144, 147, 150, 151, 152, and 153, and most of the lines made available to us through the courtesy of the U. S. Navy Hydrographic Office. Since new data are coming in continuously, it was deemed inadvisable to make a detailed presentation of the material examined so far. Therefore, except for a few detailed local surveys, we will limit ourselves to a very generalized description of the various types of submarine topography in the North Atlantic.

Plate 2 was drawn to illustrate (1) the general trend of the Ridge between Lat. 40°N. and 29°N.; (2) the distribution of the known conspicuous sea mounts; (3) the extent of the data upon which we have determined the generality of certain types of submarine topography; (4) the control upon which to a large extent our 1600-fathom contour was based. Where no lines are shown, we either omitted this contour or used the few scattered soundings available on the H. O. 0955 and 0956 charts.

The fathograms shown in Plates 3, 4, 5, and 6 were obtained during the Atlantis cruise No. 150 and illustrate the major types of topography in the North Atlantic. We will confine ourselves mostly to a description of the area covered on this cruise.

A study of the lines in Plate 2 has shown that many of the types of topography encountered are characteristic of broad areas of the ocean floor.

On cruise 150 the Atlantis went from Woods Hole to Bermuda, from Bermuda to the Azores

and back, thus crossing the northwestern basin of the Atlantic (the "North American Basin") and the Mid-Atlantic Ridge. These areas were characterized by several distinct types of bottom topography. Thus the slightly shoaler part of the North American basin, which was crossed on the trip from Woods Hole to Bermuda, has somewhat different characteristics from those of the deeper part, between Bermuda and the Ridge. The central and deepest part of the latter is occupied by a great horizontal plain at the depth of 2900 fathoms. The areas bordering this basin, on the other hand, were characterized by more rugged topography, and the eastern border was quite different from the western. The rugged and high central part of the Mid-Atlantic Ridge is quite different from its shelved flanks. The reader should realize that this subdivision of the ocean into separate zones is by no means arbitrary, but corresponds to what appear to be distinct physiographic provinces of the ocean bottom.

Ocean Bottom Between Cape Cod and Bermuda

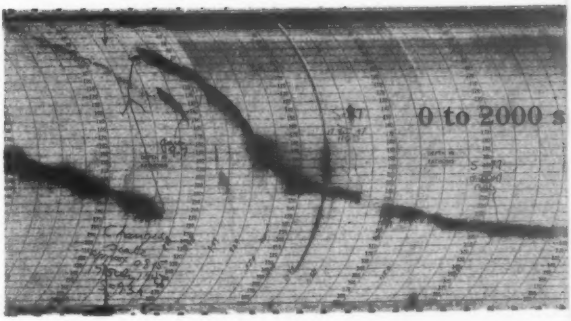
On cruise 150 one continuous fathogram from Woods Hole to Bermuda was obtained. This has since been matched by numerous other fathograms taken by the Atlantis and other ships. Portions of the fathogram of cruise 150 are reproduced in Plate 3. This shows a succession of steplike horizontal or imperceptibly sloping levels at 1500, 1950, 2100, and 2500 fathoms (Pl. 3a). The last three are more pronounced than on a typical crossing of the continental slope. However, a 1500 ± 100 level is a fairly general feature of the continental slope between Lat. 40° and 35°N. The deepest section of ocean floor along this line occurs about half way between Woods Hole and Bermuda, at 2650–2700 fathoms. It is about 250 miles wide and is remarkably flat and featureless, except for one fair-size sea mount at 36°38′N., 67°55′W. Unfortunately this sea mount was encountered on the homeward bound course, when the fathometer was not recording. Subsequent cruises by the Atlantis (151 and 153) and the R/V Caryn (winter of 1947–1948) have shown that it rises to 1590 fathoms.[1] During cruise 150 the Atlantis obtained a core 20 miles southeast of this mount. It showed coarse quartz sand stratifications in the upper section of the core. This sand is now being studied by D. Ericson. Recent (Trask, Phleger, and Stetson, 1947) investigation has shown the presence of sand in the deep basin of the Gulf of Mexico. More recent cruises of the Atlantis have found sand at depths of many feet in various parts of the North American basin. The presence of these sands poses one of the most baffling problems of submarine geology.

At Lat. 34°N. on the track of A150, at the end of the 2700-fathom stretch, the bottom starts a gentle overall rise toward Bermuda. The ocean floor on this slope is quite different from that noted off the continental shelf. The succession of levels is not so apparent as it was on the continental rise. Numerous steep sea mounts (10- to 30-degree slopes, rising between 100 and 500 fathoms above the ocean floor on which they are situated) dot the gentle overall rise (Pl. 3). Only in between these peaks is there any indication of the existence of levels. Thus Plate 3 shows a shelf at 2500 fathoms, before the steeper rise toward Bermuda starts. This shelf is even more pronounced to the east (A151). It is preceded on all available fathograms by a sharp peak.

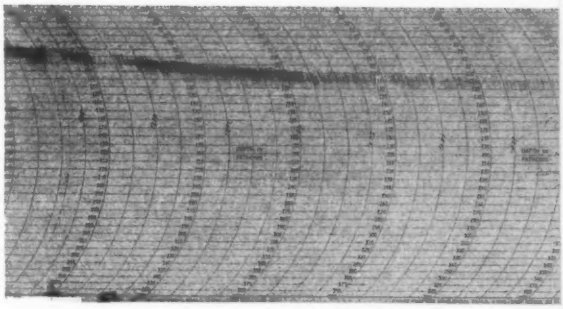
The Vicinity of Bermuda

Figure 1 shows a bathymetric map of the approaches to Bermuda, on a plotting-sheet scale, and the eight Atlantis tracks upon which these contours are based. The base of the island has a NE.–SW. trend and is about 100 miles wide in a NW.–SE. direction. Approaching Bermuda from the north (Pl. 3d), at the depth of 2450–2500 fathoms the first change in slope occurs in the slow upward grade toward the south. After this the bottom rises much more rapidly. These fathograms show the large sea mount 25 miles north of Bermuda, rising to less than 500 fathoms. The steepest part of the slope is between 850 and 100 fathoms; its average inclination is of the order of 20°–25°.

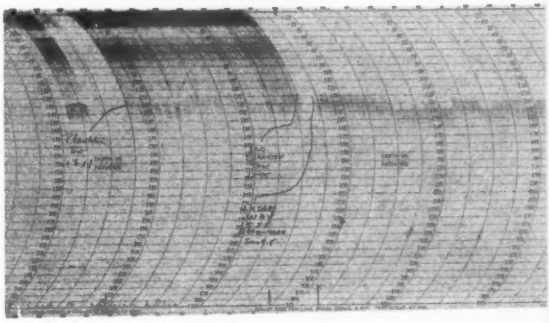
[1] Recent data show that the whole section of the North American basin northeast of the Atlantis 150 track is characterized by a flat plain at 2650 fathoms, dotted by a number of large flat-topped sea mounts some of which rise to less than 800 fathoms (Pl. 2). A number of these mounts appear to string out in a southeast direction, away from Cape Cod. A detailed report on this area is now in preparation.



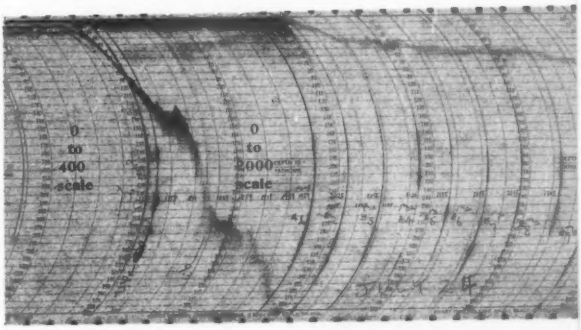
a. Continental slope. Note the pronounced almost horizontal shelves between 1450 and 1600 fathoms and between 2050 and 2100 fathoms.



b. The flat horizontal bottom between 2600 and 2700 fathoms. The irregularities of this track are due to frequency changes.



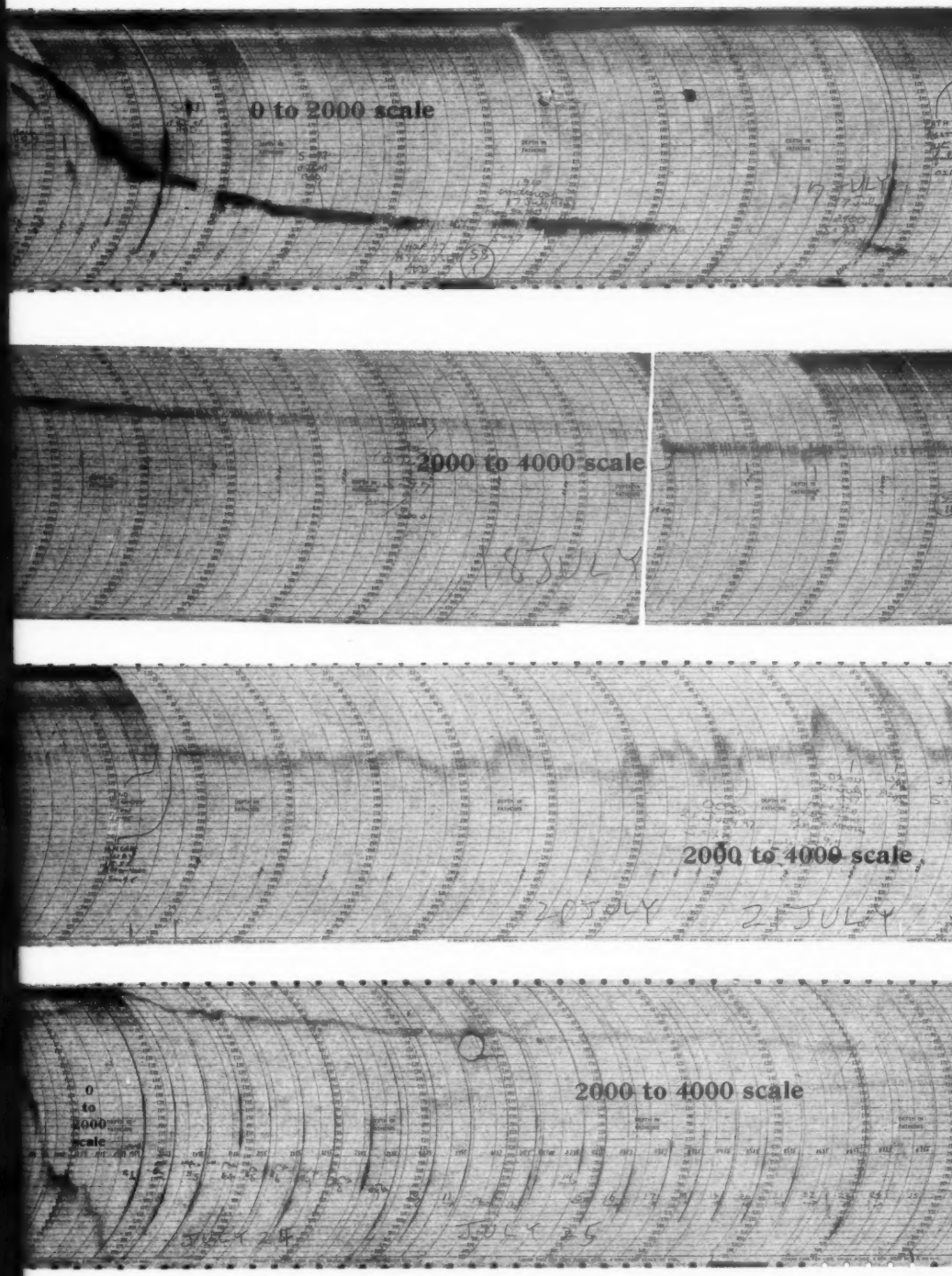
c. The bottom sloping up toward Bermuda from the northwest.



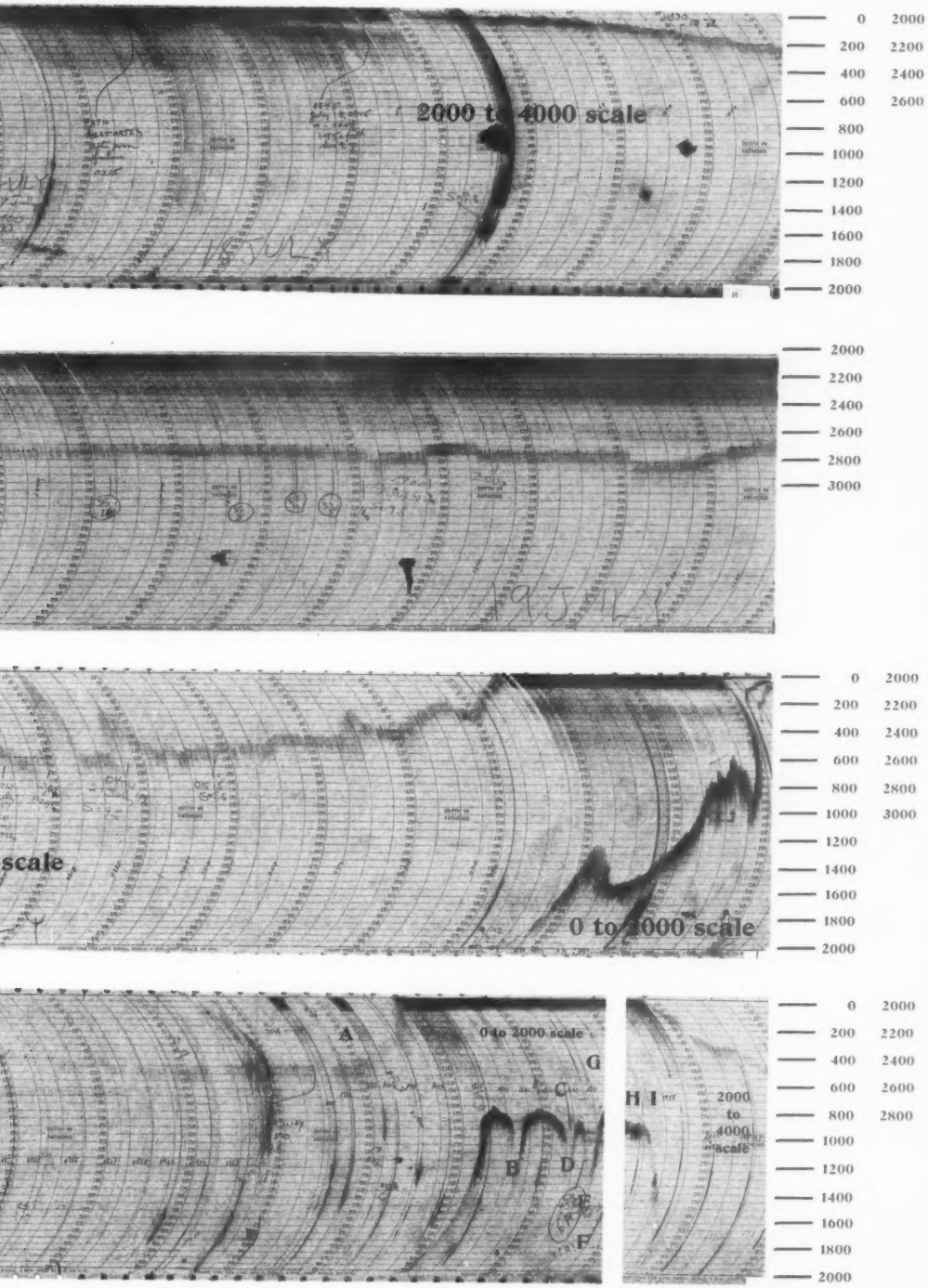
d. The northeastern slope of Bermuda. The sea mount at the extreme right is the one shown in Figure 2 with the same letters corresponding to course changes.

SECTIONS OF FATHOGRAMS

The portions of track corresponding to these fathograms are represented by the heavy lines c
(figures at right or left extremities of plates) are in fathoms, the column to the left indicating

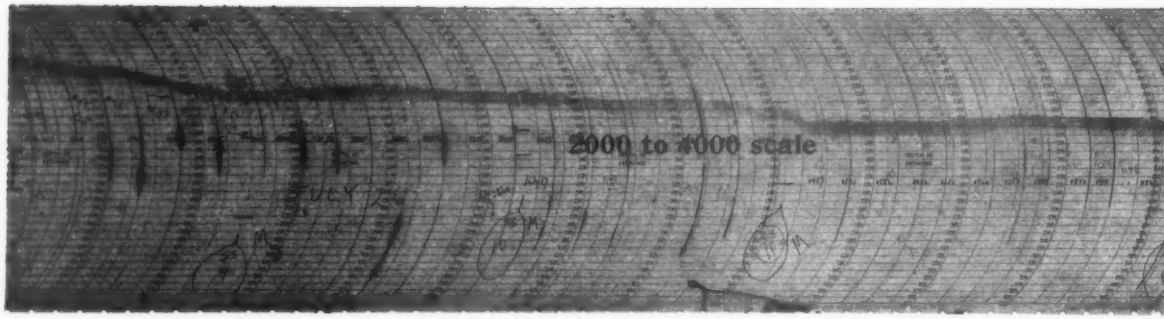


SECTIONS OF FATHOGRAMS OBTAINED BY THE ATLANTIS IN THE AREA OF BERMUDA AND
hese fathograms are represented by the heavy lines on Plate 2. Each fathogram represents between 250 and 300 miles of ocean bottom.
tes) are in fathoms, the column to the left indicating depths on the 0 to 2000 scale, the column to the right giving depths on the 2000 t

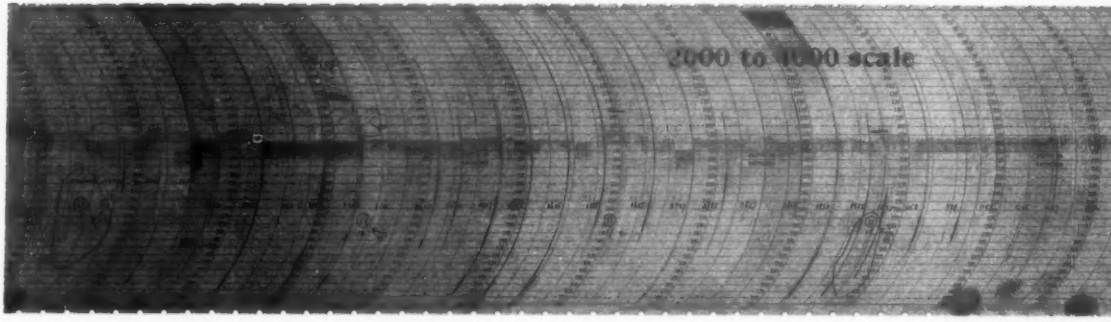


MUDA AND BETWEEN BERMUDA AND WOODS HOLE

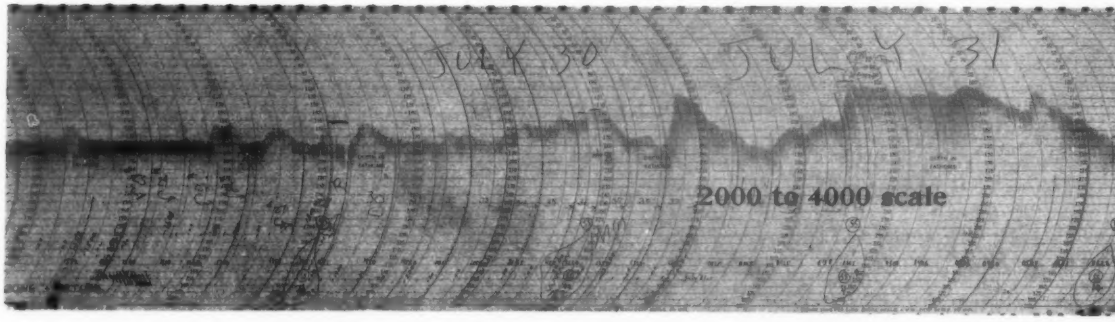
ocean bottom. Each horizontal division represents from 2 to 3 miles of track, depending on the ship's velocity. The depths on the 2000 to 4000 scale.



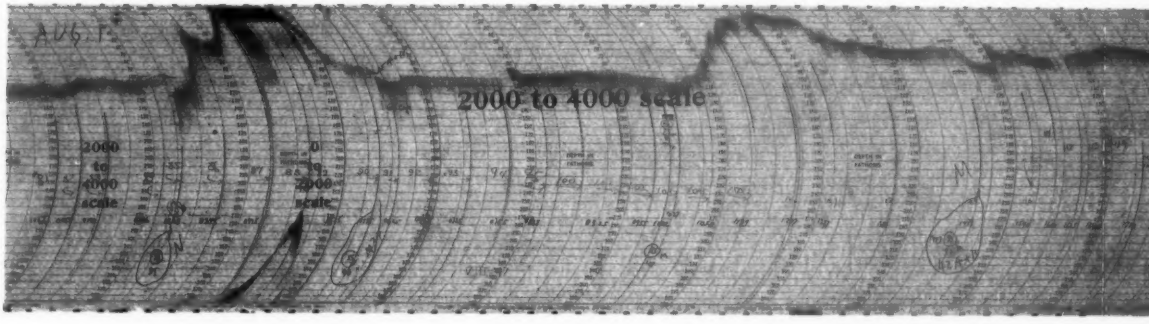
a. Portion of the bottom sloping toward the 2900-fathom plain.



b. Western portion of the 2900-fathom plain.

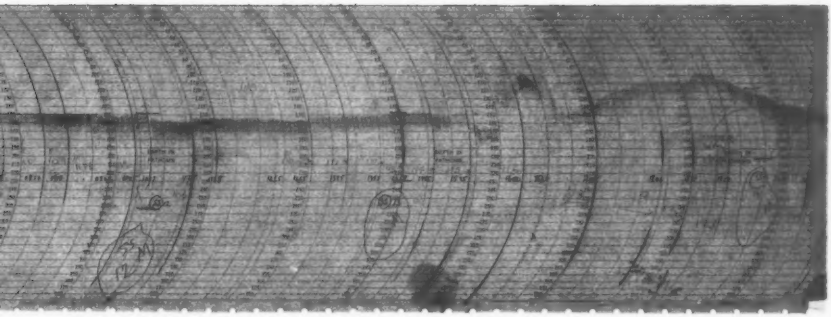


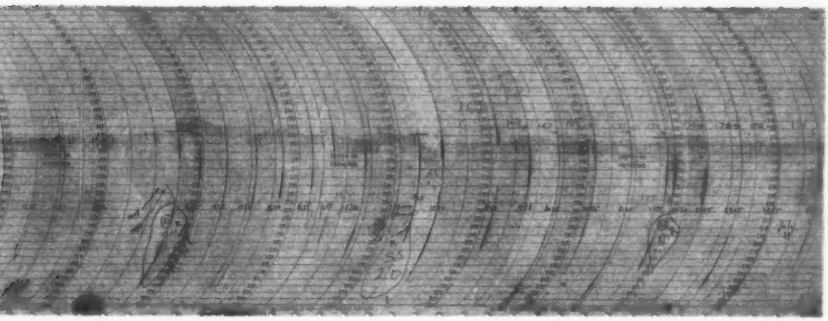
c. Eastern end of the 2900-fathom plain and the mountainous type of topography which characterizes the ocean bottom between the 2900-fathom plain and the Terraced Zone of the Mid-Atlantic Ridge

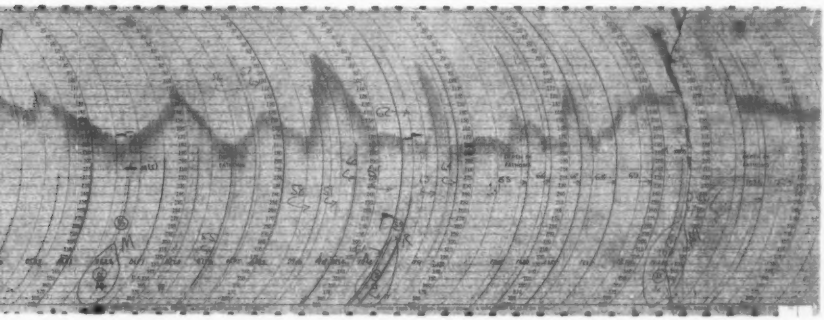


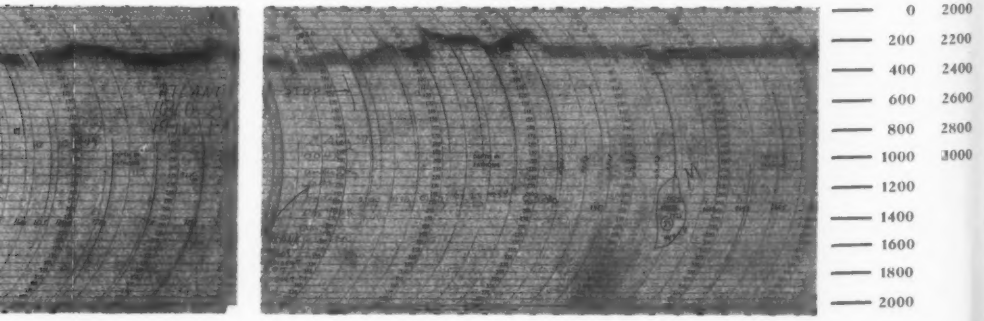
d. Beginning of the Terraced Zone and the large sea mount preceding the 2500-fathom terrace, on the western flank of the Ridge.

FATHOGRAMS ILLUSTRATING THE 2900-FATHOM PLAIN AND THE 250









AND THE 2500-FATHOM TERRACE

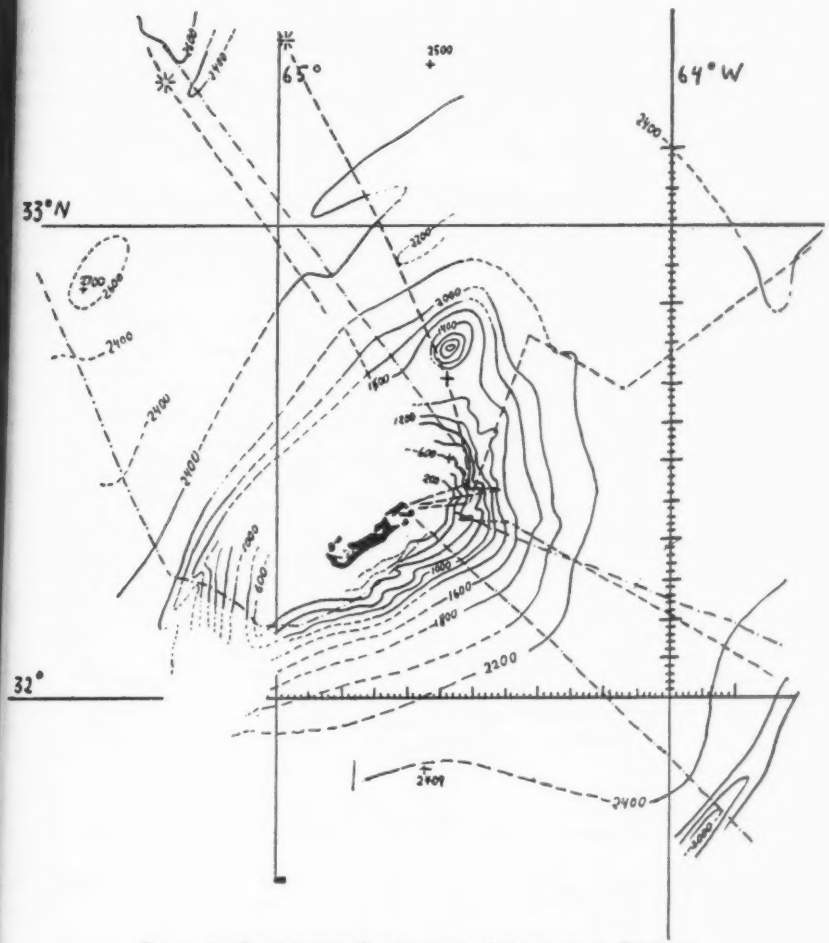


FIGURE 1.—BATHYMETRIC CHART OF THE APPROACHES TO BERMUDA
Contour interval 200 fathoms.

Similarly, as one goes east from Bermuda the bottom drops off very rapidly to 2200 fathoms, and then more slowly to 2500. This seems to be the depth at which, as far as all the north, northeast, east, and southeastern approaches toe Bermuda are concerned, the bottom may be said to start feeling the influence of the island. On the northeastern course from Bermuda the 2500-2550 fathom stretch of sea floor extends for about 32 miles. It is then followed by a rise, gentle at first, then steeper and steeper until the depth of 780 fathoms is reached 130 miles from Bermuda. This sea mount was originally discovered by the U.S.S. MUIR on its cruise from Philadelphia to Dakar in 1945. The writers conducted a box survey here (Fig. 2); this showed that the sea mount has a flat top at about 780 fathoms. It is elongated in a NW.–SE. direction. Its exact length is unknown, but it is probably less than 30 miles.

The fathogram is reproduced in Plate 3d. Its appearance recalls somewhat the description given by Hess (1946) of the Pacific guyots, and the sea mounts of the Gulf of Alaska described by Murray (1945). A short core obtained from the summit shows a discontinuity between a

13-cm layer of Recent sediment and an underlying chalky mud containing Eocene Foraminifera, suggesting that in recent times some scouring agent has been active close to the surface

miles wide. Thence a generally level plateau at 2400 fathoms extends for about 60 miles, followed by a steplike succession of flat stretches at 2550–2600 fathoms and 2750–2800 fathoms,

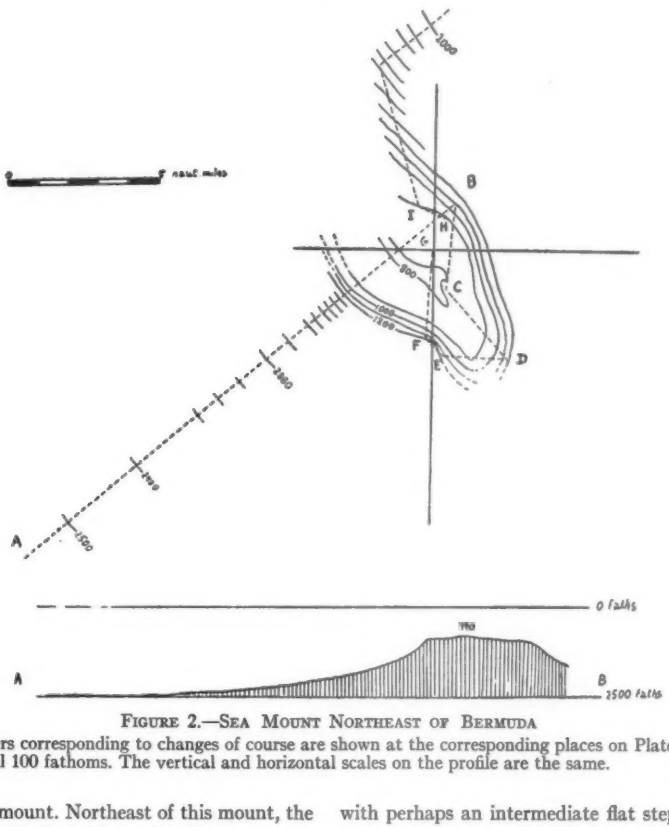


FIGURE 2.—SEA MOUNT NORTHEAST OF BERMUDA

The letters corresponding to changes of course are shown at the corresponding places on Plate 3d. Contour interval 100 fathoms. The vertical and horizontal scales on the profile are the same.

of this sea mount. Northeast of this mount, the bottom drops off rapidly to 2700 fathoms.

THE OCEAN FLOOR BETWEEN BERMUDA AND THE 2900-FATHOM PLAIN

The section of ocean floor here considered extends from points A to B of Plate 2 on the northern A150 track.

The fathogram reproduced in Plate 4a shows gentle topography in this section of the ocean floor. After a short (3 miles) 2700-fathom stretch northeast of the sea mount just described, it is characterized essentially by a long gentle swell which rises to 2300 fathoms and is about 60 with perhaps an intermediate flat step at 2650 fathoms having a width of 20 miles (Pl. 4a).

The end of the 2750 level seems to be characterized by a narrow (2 miles) flat-bottomed trough (2840 fathoms) followed by a sharp peak (2350 fathoms) and a broad hump rising to 2560 fathoms and 20 miles wide (Pl. 4a).

To the south the rise from the 2900-fathom plain to Bermuda is more rugged. Thus all soundings pertaining to this area show that there are numerous peaks, separated by deep valleys.

The fathograms obtained between Bermuda and the 2900-fathom plain confirm this. They show that the bottom consists of a succession

of much sharper peaks, separated by deep valleys and narrow flat-bottomed trenches reaching much greater depths than any valley on the northern track (2900 to 3100 fathoms). Judging from existing data, their trend appears to be northeast-southwest.

The 2900-fathom Plain

At point B on the northern route (Pl. 2) a depth of 2900–2880 fathoms is reached. At this point and depth begins what is perhaps the most remarkably flat stretch of ocean floor in the Atlantic. It extends, with a few knolls here and there, for over 300 miles of track (B to C on Pl. 2).

A typical fathogram obtained in this area is reproduced in Plate 4b. Both the western (Pl. 4b) and the eastern extremities (Pl. 4c) of this plain are characterized by a multiplication of the otherwise small number of knolls. These look like small pimples interrupting the otherwise horizontal ocean floor. This 2900-fathom plain is now known to occupy a very large section of the North American basin. Though we cannot yet delineate its boundaries precisely, owing to the insufficient amount of continuous fathograms, it may extend as far north as Lat. 40°N. In these latitudes it might be 50 or 100 fathoms shoaler. Several fathograms show it to be several hundred miles wide in the vicinity of Lat. 31°N., though it appears to be marked by more numerous and higher knolls. Many of these rise to over 100 fathoms above the level plain. Also several higher peaks rise 1000 fathoms or more above this level. Some of these appear to be ridges having either due north or east-northeast trends. The level plain is somewhat deeper (2950 to 3000 fathoms). Its beginning, at 31°N., 58°W. on the fathograms of A151 is characterized by a high peak (about 1000 fathoms in height), having slopes of the order of 20° to 30° on the fathogram. (The width of its base is about 9 miles.) In other words, when the slope correction is considered, its walls would perhaps be more than 40° steep (Fig. 3). It is preceded by a small trough, 2–3 miles wide, be the bottom of which is horizontal at 3100 fathoms.

The fathograms of Atlantis cruise 151 show a similar plain occupying the floor of the basin east of the Ridge, between the Ridge and the northwestern coast of Africa (the northern Canary basin). It appears therefore that this 2900-fathom plain is of great regional significance. Its aspect on the fathograms suggests that vast portions of the ocean floor have remained tectonically undisturbed since their formation. The small knolls may be due to one of several causes: block faulting, lava escaping through fissures, or submarine volcanoes. Their true shapes and distribution pattern over the ocean floor remain to be ascertained.

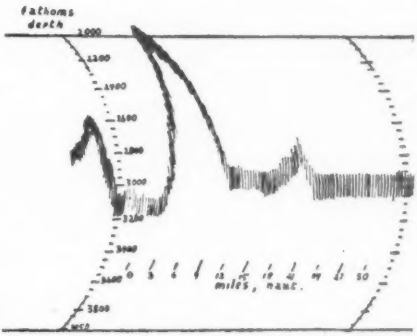


Figure 3.—Sea Mount at the Western Edge of the 2900-Fathom Plain at Lat. 31°N.

This is a tracing of a fathogram showing a sea mount located at the point marked X on Plate 2. (31°N., 58°W., approximately).

The Ocean Floor between the Eastern Edge of the 2900-fathom Plain and the Foot of the 2500-fathom Terrace

The 2900-fathom plain ends rather aburptly at point E on Plate 2 against a series of high peaks, one or two of which have slopes exceeding 30°, and heights of as much as 500 fathoms (Pl. 4c). All available fathograms between Lat. 36° and 30°N. show the presence of this mountainous region between the 2900–3000 fathom plain and the "Terraced Zone" of the Ridge. Seismic work by Ewing and Press aboard the Atlantis on cruise 150 has shown that this area gives a different type of reflection, indicating the presence of sizable amounts of mud (several hundred feet), as against practically none on the 2900-fathom plain and much larger quantities in the Terraced Zone of the Ridge.

A sea mount of exceptional dimensions is located at 35°12′N., 45°33′W. Its western foot is marked by a small trough (2600 fathoms). It rises from this depth to 1300 fathoms (Pl. 4d). It marks the end of the mountainous area and the beginning of the Terraced Zone of the Ridge.

The Terraced Zone

On the eastern side of the peak we have just described the bottom drops off to 2500 fathoms, at which depth it remains approximately flat for about 35 miles (Pl. 4d). On the track of Atlantis 150, along Lat. 31°N., a 2500-fathom level stretch of the same width was found, preceded also by a high peak at approximately 31°N., 45°30′W. shoaling to 1800 fathoms. One cannot say what relation these two remarkably similar structures bear to one another.

This marks the beginning of the succession of flat stretches characterizing the lower flanks of the Ridge, which, for lack of a better term, we shall call terraces. The area characterized by this type of topography we shall henceforth call the Terraced Zone. The authors do not wish to imply anything concerning the origin of these flat stretches by the use of the name "terrace." It was merely deemed more advisable to use some familiar term instead of coining a new name.

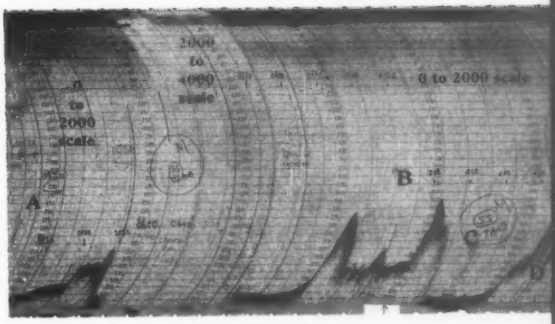
Along the Atlantis 150 track the 2500-fathom terrace ends against a 2000-fathom peak bordering a 2300–2350-fathom terrace, approximately 35 miles wide (Pl. 4d). The inner eastern edge of this terrace is marked by a broad hump of small elevation, itself broken up into a series of narrow ridges shoaling to 2100 fathoms. Next comes a 2250–2270-fathom terrace 25 miles wide and characterized by its remarkable flatness. Similar terraces appear on all fathograms approaching the Ridge between Lat. 35° and 30°N. at approximately the same depths. Conceivably the differences in depth for terraces found along parallel lines would not always be real, but would be due in part to instrumental causes.

The 2250-fathom terrace marks the beginning of that portion of the Ridge and its flanks which was closely investigated during the summer of 1947. The zone between the 2250 and 1600 isobaths is 60–100 miles wide on both sides of the Ridge. It is characterized by a discrete series of flat terraces that occur at the same depths on both sides of the Ridge, at points hundreds of miles apart. The reflection shooting studies done during the same cruise by Ewing and Press indicate great thicknesses of sediments on these terraces (up to 1000 feet and more). The fathograms reveal that the most conspicuous terraces, remarkable both in numbers, in different localities, and in total length, are at depths of 2250, 2150, 2100, 2050, 1950, 1850, 1800, 1750, 1700, and 1600 fathoms. Of these the most numerous were at 2250, 2050, 1950, 1850, and 1800 fathoms. Plate 5 reproduces a few typical records from the terraced zone. We have given no detailed charts of the exact distribution of all the individual terraces enumerated above because the number of tracks available over most of the terraced zone is insufficient as yet to enable us to establish any definite correlations between most of them. However, in a few cases it has been possible to establish without much doubt the continuity of some of these terraces over great distances.

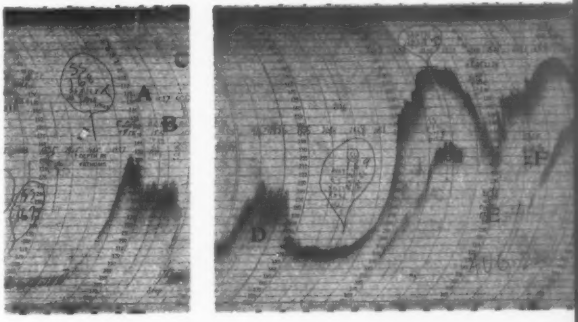
In Plate 7 the 1900-fathom terrace might be continuous for more than 90 miles and appears to follow closely the general NE.–SW. trend of the Ridge. The 2050-fathom terrace might be continuous for about 80 miles. The corresponding fathograms are reproduced in Plate 5a. The fathograms of the crossing of the Atlantis 153 cruise over this area show that the 2100-fathom level (Pl. 7) extends for perhaps another 100 miles farther west. Possibly, therefore, these terraces form a series of horizontal fringes contouring the outer edge of the higher portions of the Mid-Atlantic Ridge.

Plate 5b shows the fathogram obtained in the region of point D on Plate 2. The summit of the peak on the extreme right corresponds to the change in course made at that point, indicating that the change occurred before reaching the true summit of the sea mount. It emphasizes the possibility of a correlation between the 1700-, 1750-, and 1850-fathom terraces on each side of the sea mount. The portions of these visible on the fathogram would be sections of what may well be a continuous series of terraces circling the flanks of the sea mount.

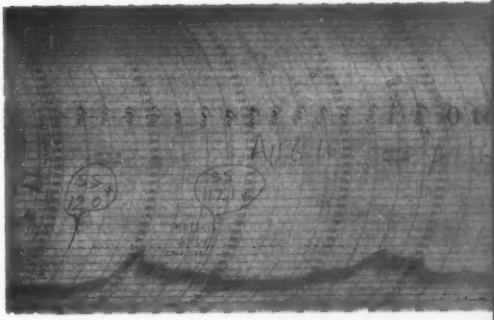
Many of the terraces described are horizontal, but in some there is evidence of tilting. This is particularly well illustrated by the fathogram



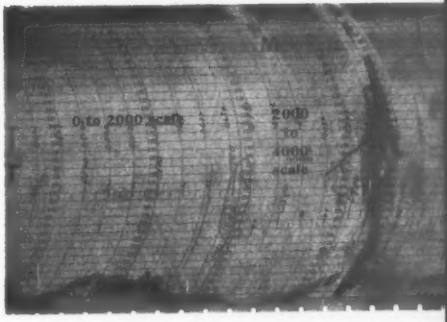
a. A section of the Terraced Zone represented with the same letters for course changes on Plate 7. Western flank of the Ridge.



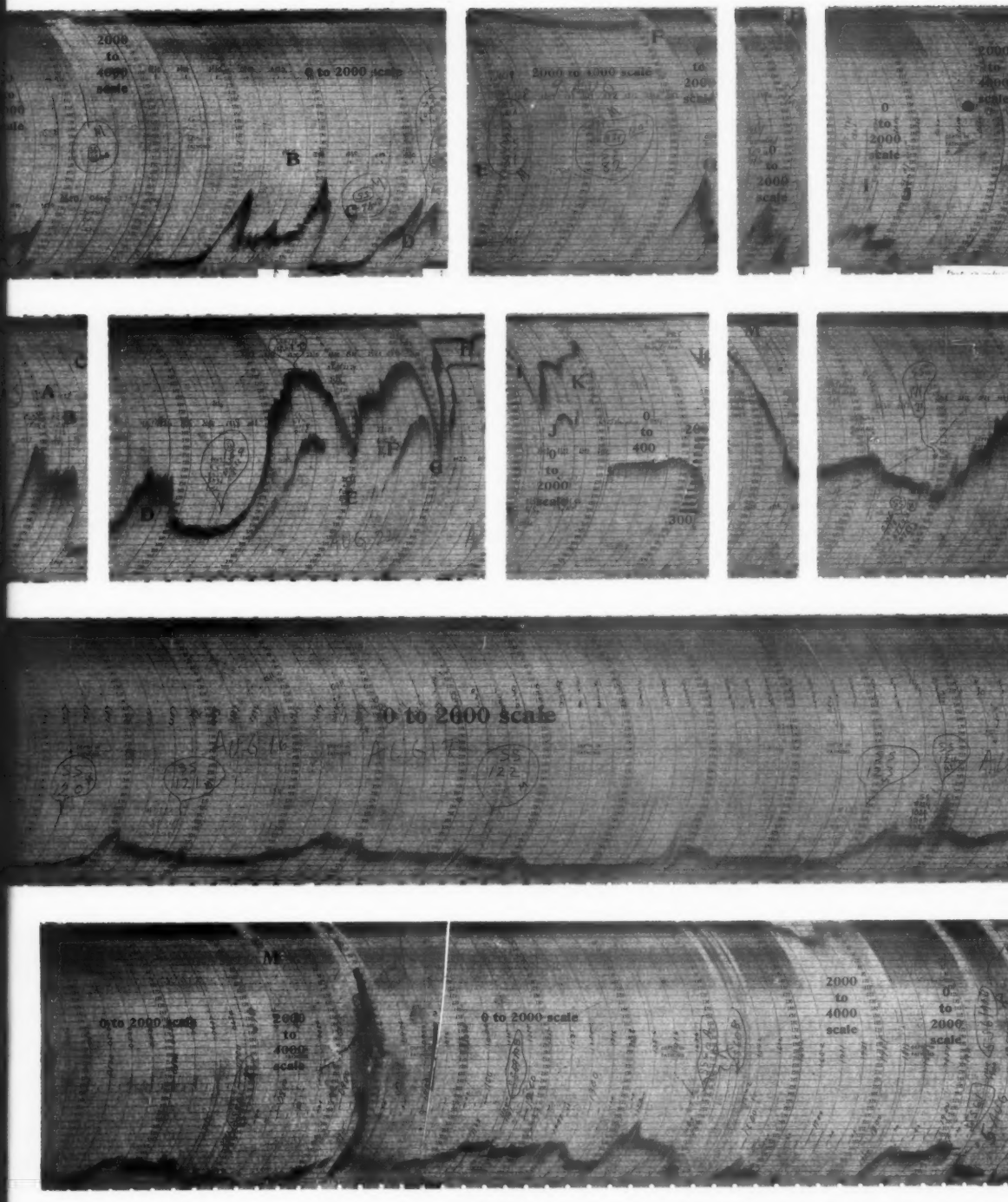
b. Terraced Zone southwest of the Azores, with the large sea mount represented in Figure 4. Eastern flank of the Ridge.



c. Terraced Zone southwest of the Azores. To the right, the large peak with the steep southwestern slope marks the beginning of the Azores plateau. Eastern flank of the Ridge.



d. Terraced Zone on eastern flank of the Ridge immediately south of area on Plate 7. The narrow trench at point M (*see* Plate 7) is interpreted as the prolongation of the East-West trough shown in Plates 7 and 8.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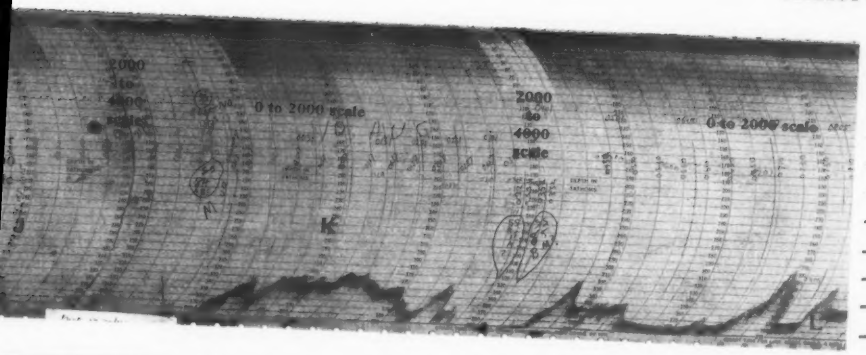

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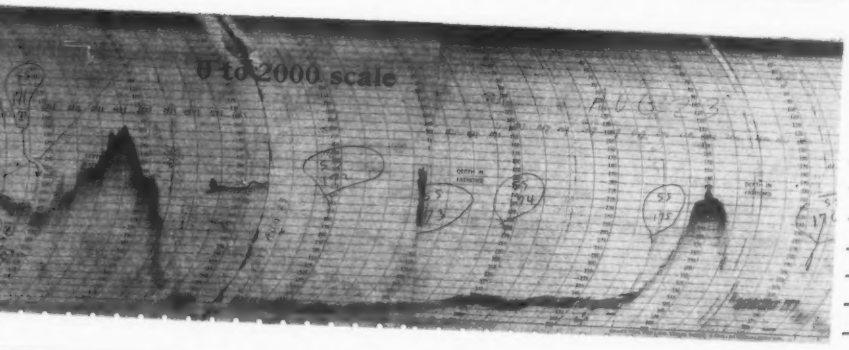

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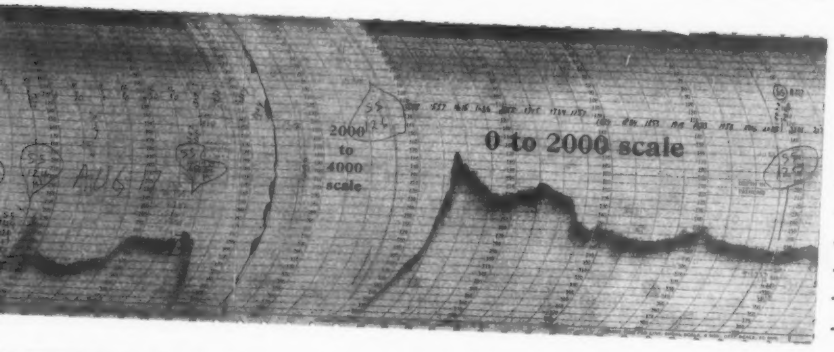

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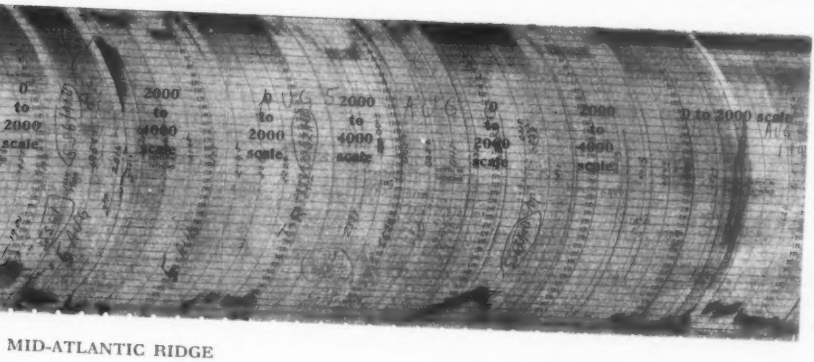

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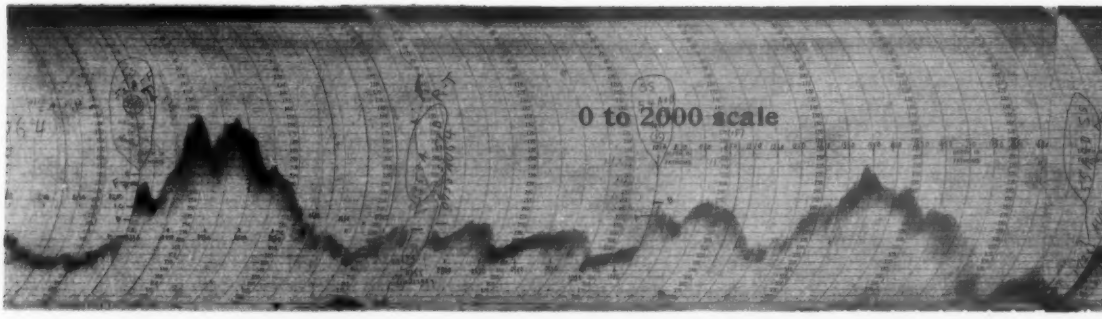
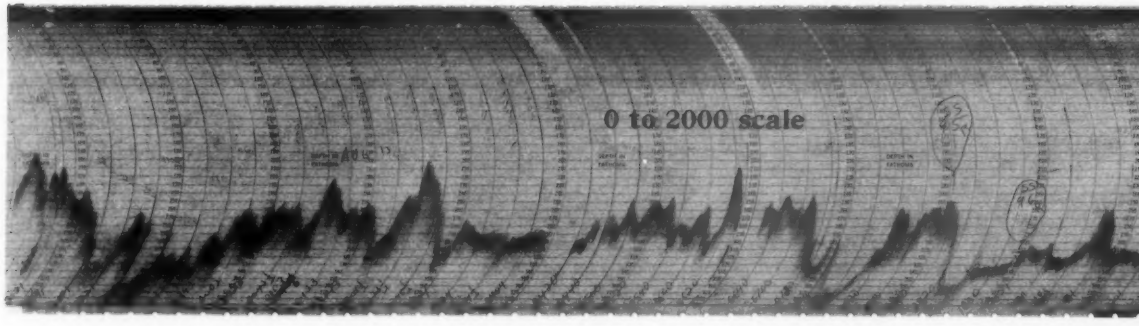
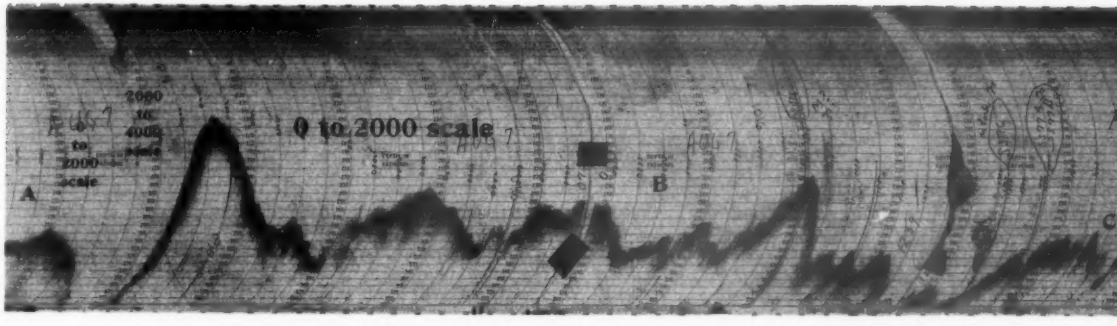
MID-ATLANTIC RIDGE

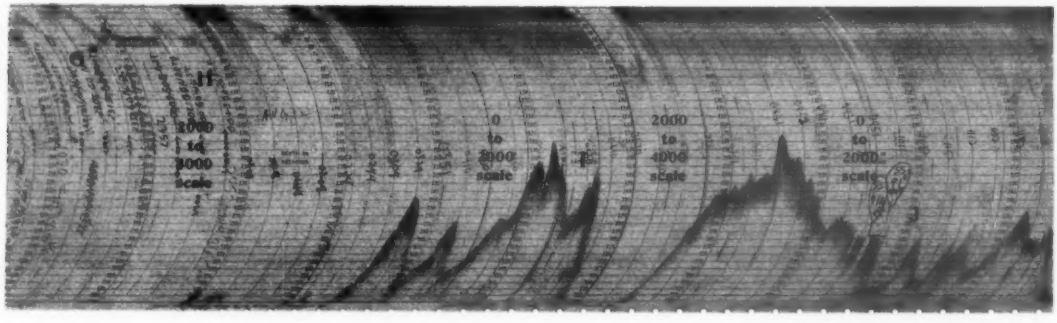


a. and b. Sections of the Main Range. Note that the right-hand side of b shows the transition to the Terraced Zone, on the eastern flank of the Ridge.

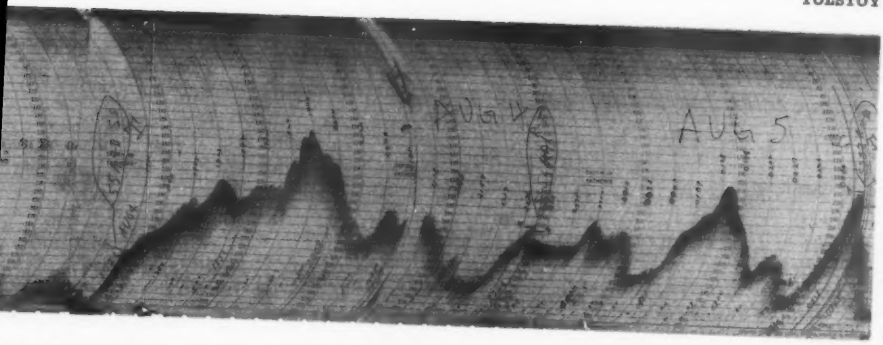




c. Two tracks across the trough of Plate 8, with some typical Main Range topography between. Letters correspond to Plate 8.



d. Two tracks across the same trough; letters correspond to Plate 8.



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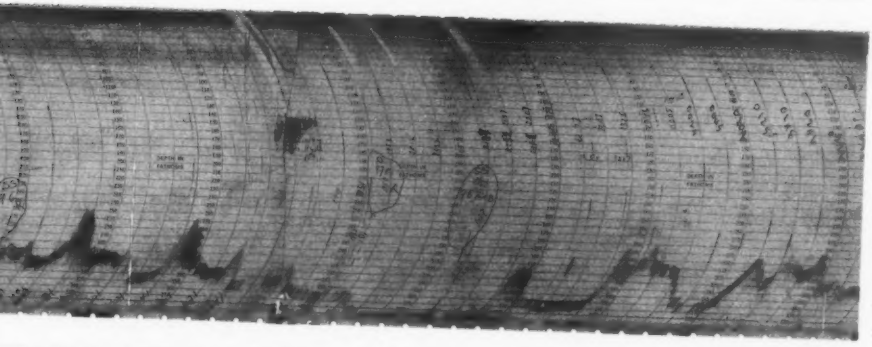

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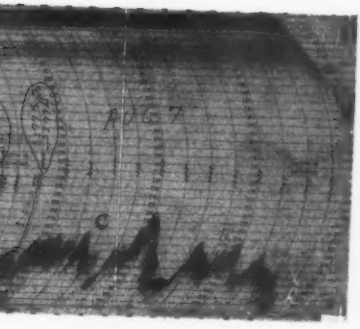

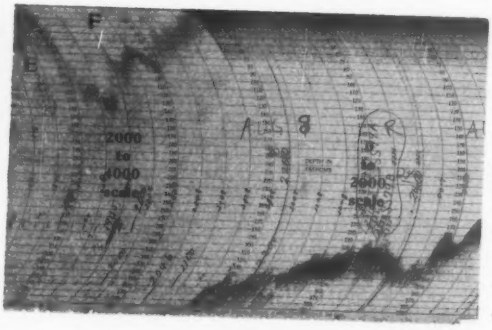

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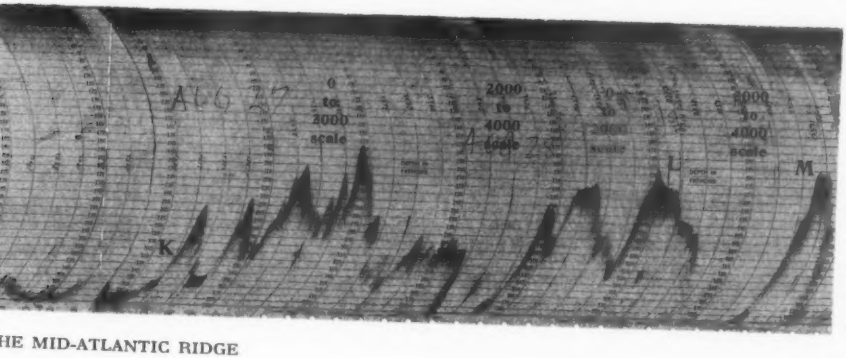

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shown in Plate 5c in which a break is suggested in the center of a 1750-fathom terrace with subsequent tilting of the left (southwestern) portion. Many of these terraces are bordered on their outer edge by peaks from 50 to several hundred fathoms in height. Others are enclosed on both sides by steep walls having slopes of as much as 30° to 35°. Examination of the fathograms obtained by the ATLANTIS cruise 151 reveals that the same terraces, at exactly the same depths and characterized by the same horizontality over many miles, are to be found between 1600 and 2500 fathoms off the coast of North West Africa. Plate 5d shows the aspect of the terraced zone immediately south of the area represented by Plate 7.

THE MAIN RANGE OF THE MID-ATLANTIC RIDGE

This part of the Ridge is notable for the absence of terracing. It consists essentially of a series of parallel ridges trending NE.–SW., and shoaling in places to less than 800 fathoms. Plates 6a and b show a few samples of the records obtained over the Main Range.

Only at the outward edge of the Main Range is any evidence of terracing found; it is in the form of flat-bottomed valleys (Pl. 6a, b; Pl. 5a) that appear mostly in the transition zone between the Main Range and the Terraced Zone The depths of their level floors are those of the shoaler series of terraces recognized elsewhere in the terraced zone: 1600, 1700, 1750, 1800, and 1850 fathoms.

Wherever box surveys were conducted over the Main Range or where the intersection of two of the ship's courses made the drawing of isobaths possible, the ridges show a remarkably constant NE.–SW. trend. This is also the trend of the Main Range as a whole in the area surveyed, as may be seen from Plate 2. Plate 7 emphasizes the NE.–SW. trends of the individual ridges of the Main Range.

Dredging the eastern flank of the steep ridge running along the northwest edge of the 2100-fathom terrace at point G of Plate 7 brought up a large block of manganese-encrusted limestone of probably Cenozoic age.

A major exception to the NE.–SW. trends of this area is a deep trough running E.–W. along Lat. 30°N. from 41°W. to 43°W., which can be seen in the southwestern section of Plate 7. A detailed survey was performed in this area (Pl. 8; Pl. 6c, d). The trough is bounded by steep walls having in places average slopes of the order of 30°. The deepest point of the trough is at 2800 fathoms in the western part. The bottom appears to be irregular; two deeper sections (more than 2600 fathoms) are separated by a narrow pass at 2200 fathoms. Several dredging stations were made here, on both the northern and southern flanks of the trough, in the general area of the end points of the tracks. In each case numerous rocks were brought to the surface. Shand (1949) identified them as being strongly or partly metamorphosed ultrabasics: peridotites, olivine gabbro, and olivine basalt. Many of them show evidence of crushing and slickensiding. One or two specimens are practically fault breccia. Some of the basalt fragments show a glassy surface, indicating rapid chilling. Both the topography of this trough and the nature of the rocks obtained therefrom suggest that it may be a graben. It seems significant that this feature is in the center of one of the seismically more active regions of the Mid-Atlantic Ridge (between Lat. 25°N. and 35°N.). This and the absence of traces of the terraced type of topography along the walls of this trough would indicate that it is a fairly recent feature, superimposed upon the typical Mid-Atlantic Ridge type of topography.

Finally we call attention to the remarkably large sea mount at point E on Plate 2. (*See also* Figure 4 and Plate 6c, left end.) Here is another sea mount of the flat-topped variety, which rises to less than 200 fathoms from the surrounding bottom at 1900 fathoms. Dredging the top yielded a large number of calcareous discs characterized by peculiar cavities in their centers. Whereas the outside of the pebbles is relatively smooth, the interior of these cavities is rough. Their dimensions are variable; they average about 15 cm long and about 4 cm thick. Thin sections of these discs reveal the presence of numerous pteropods. According to D. Ericson, all are species that could have inhabited the ocean from the early Tertiary to the present. None are sufficiently characteristic of a specific period to date this limestone.

The fathograms taken in the vicinity of the

Azores show that the plateau on which these islands are situated is limited on its southwest side by a steep slope about 600 fathoms high (Pl. 6c). The topographic map which resulted from the investigations of the ALTAIR shows that the Azores plateau is limited on its sides by such a slope and forms an easterly bulge of the Mid-Atlantic Ridge.

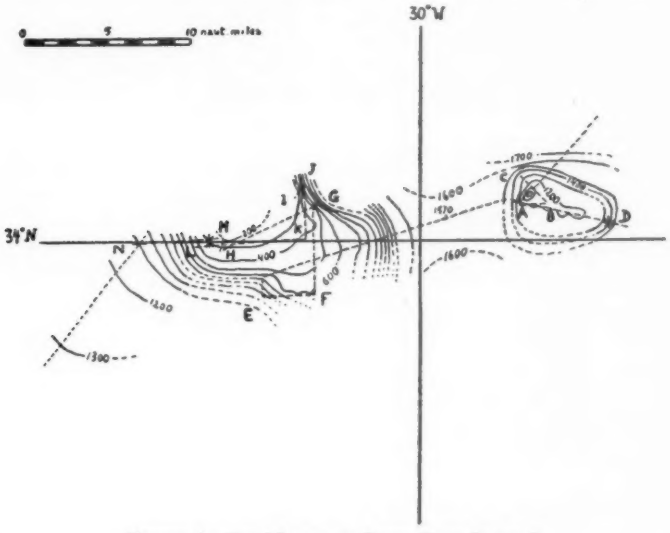


FIGURE 4.—SEA MOUNT AT POINT E OF PLATE 2

This is a sketch-map interpretation of the fathograms shown in Plate 6c of the flat-topped sea mount southwest of the Azores. The letters correspond to those of Plate 6c. Contour interval 100 fathoms.

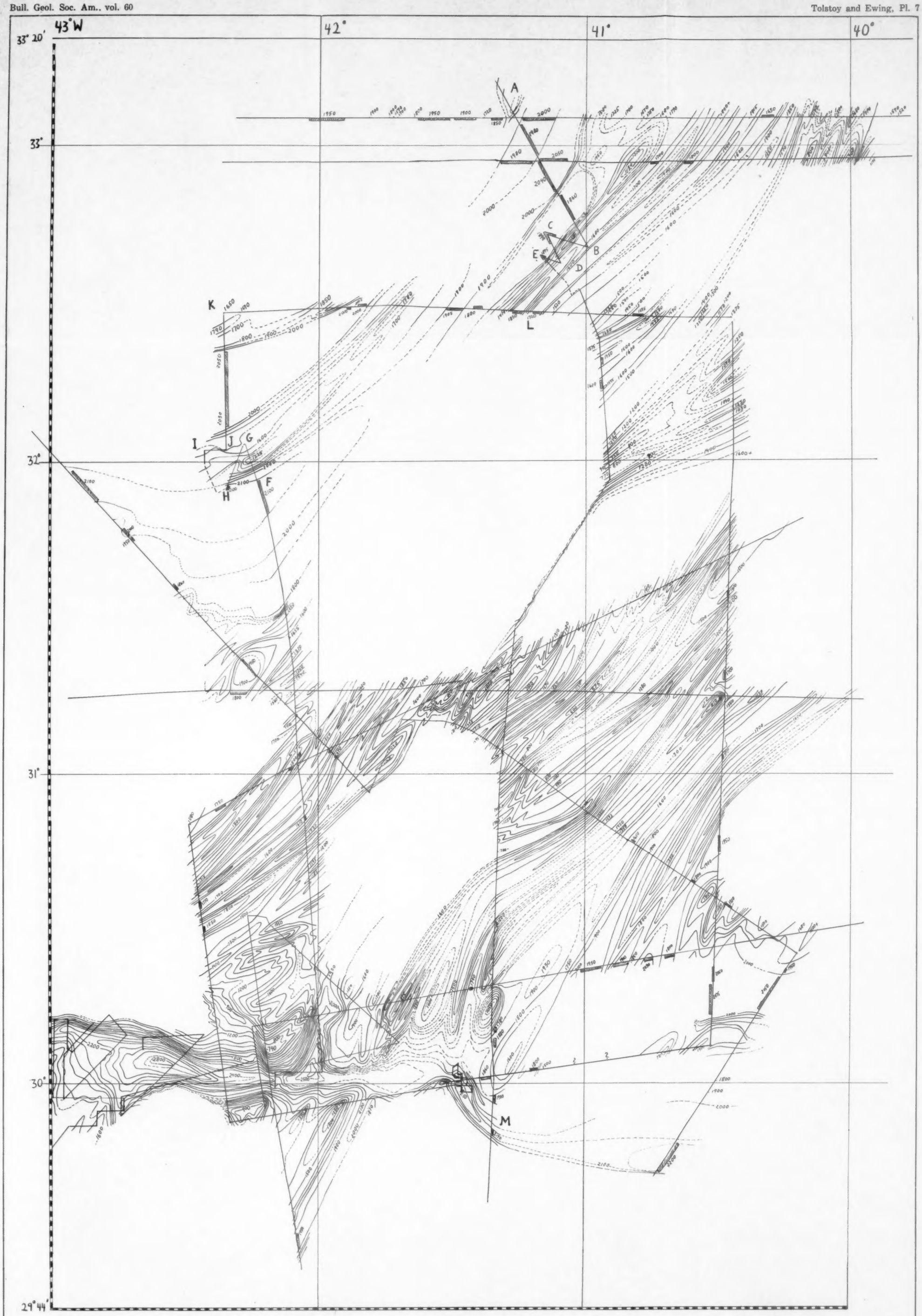
The type of topography shown on the fathograms on the Azores plateau obtained by the ATLANTIS was quite different from any noted on the rest of the Ridge. The surface of the plateau is characterized by a series of ridges, which the ALTAIR expedition plotted as trending NW.–SE. (*i.e.*, approximately at right angles to the trends noted in the Main Range farther south), and whose spacing is entirely different from that of the ridges forming the Main Range of the Mid-Atlantic Ridge. Instead of being close together with only a narrow steep-walled valley separating each two ridges, the ridges on the Azores plateau are separated by broad valleys. It seems improbable that two such different types of structure could have been built by the same process.

SEISMIC EVIDENCE

Reflection-shooting studies of the ocean bottom by Ewing and Press on board the ATLANTIS during the summer of 1947 indicate a greater thickness of sediment in the terraced zone than on the 2900-fathom plain west of the Ridge or on the Main Range. The latter seems to be devoid of any sediment covering the mountain slopes. Each time a mountain was approached, the sediment covering the terraces thinned conspicuously.

INADEQUACY OF DATA OBTAINED BY SOUNDINGS WITHOUT CONTINUOUS FATHOGRAMS

Until now the horizontal plains which, judging by the results of the most recent data, seem to be characteristic of the deeper section of the floor of the North Atlantic Ocean, had gone unnoticed. These sections of the ocean floor were known to be more or less horizontal, without any conspicuous sea mounts. But their unusual flatness, interrupted here and there by a few knolls, which makes them so remarkable,



SKETCH MAP OF THE MAIN AREA SURVEYED ON THE RIDGE BY THE ATLANTIS

The shaded sections of track indicate terraces

Contour interval, 100 fathoms

had escaped the notice of all previous observers. Until quite recently continuously recording fathometers had not been used in the deep ocean. All soundings pertaining to this and other parts of the ocean were made by a visual dial arrangement, and the result was a series of discrete soundings several miles apart; each sounding was probably accurate only to within ± 30 to 50 fathoms. The profile of the ocean bottom obtained from these was like an interpolated curve, with a possible cumulative error between two successive soundings of the order of 50 to 100 fathoms. It is therefore not surprising that these flat stretches of ocean floor, great portions of which are horizontal within the limits of detection by the human eye on a fathometer recording—*i.e.*, to within ±5 fathoms—, remained unsuspected for so long. With the ever more extensive use of continuously recording fathometers, such flat plains are being discovered in the Pacific Ocean also. The late H. W. Murray of the U. S. Coast and Geodetic Survey had given us the following verbal communication:

"In the Gulf of Alaska there is a flat plain lying 140 miles offshore from Yaquina head. It has a length of 200 miles in a northwest-southwest direction and lies at a depth of 1461 fathoms at its northwest end and at 1668 fathoms at its southeast end. It seems to slope or tilt toward the continental slope, where it lies in water 200 fathoms deeper."

Conclusions

The Mid-Atlantic Ridge is characterized by two very distinct types of topography:

(1) The Terraced Zone, a succession of horizontal flat stretches occurring at depths ranging from 2500 fathoms to 1600 fathoms, is found along both flanks of the Ridge, at the same depths. From the few locations that could be correlated, it appears probable that they form fringes along the foot of the Main Range. They are covered by thick sediments. These terraces, at the same depths, recur off the coast of northwest Africa. In a few cases they exhibit evidence of tilting.

(2) The Main Range of the Mid-Atlantic Ridge, above 1600 fathoms, is a succession of parallel mountain ridges, trending NE.–SW. There are no terraces in this section of the Ridge, although the higher terraces are found in valleys, the floors of which are below 1600 fathoms.

Of all the hypotheses on the origin of sections of the Mid-Atlantic Ridge perhaps the least speculative are those proposed by Hans Cloos and J. Agostinho for the origin of the type of topography on the Azores plateau. Though these hypotheses appear reasonable enough for the specific case of the Azores, it does not appear to us to apply necessarily to the rest of the Ridge, for, as we have already noted, the Azores plateau appears to be a feature distinct from the rest of the Ridge, characterized by its own particular type of submarine topography.

Apart from a few highly interesting but scanty geological facts, the evidence about the nature of the Ridge obtained thus far by us is mostly topographic, and even this leaves much to be desired. We therefore cannot conclude much about the origin of the Mid-Atlantic Ridge.

One of the big surprises of the 1947 Atlantis expedition (cruise 150) was the proof of the existence of consolidated Cenozoic limestone on the Mid-Atlantic Ridge. Possibly considerable areas of the Ridge consist of sedimentary rocks. If so, the hypothesis of folded ranges in certain parts of the Ridge should not be rejected arbitrarily, on the sole basis that it is difficult in the light of our scanty knowledge of the history of the earth's crust to explain the why and how of such a process. It must be disproved by experimental and observational data; until then it remains a possibility.

Obviously many more data in the form of hydrographic surveys and fathograms are needed. An accurate bathymetric chart of the Atlantic Ocean would contribute greatly to our knowledge of the structural history of the Earth. Such a chart would also be useful for navigational purposes, since the Atlantic sea floor is marked by numerous features which would provide excellent "landmarks".

Such a project would be greatly facilitated by a more extensive use of continuously recording fathometers. Formation surveying—*i.e.*, surveying by means of a flotilla of ships keeping radar formation and equipped with

continuously recording fathometers—would enable this work to be carried out speedily and efficiently.

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Columbia University, New York, N. Y.

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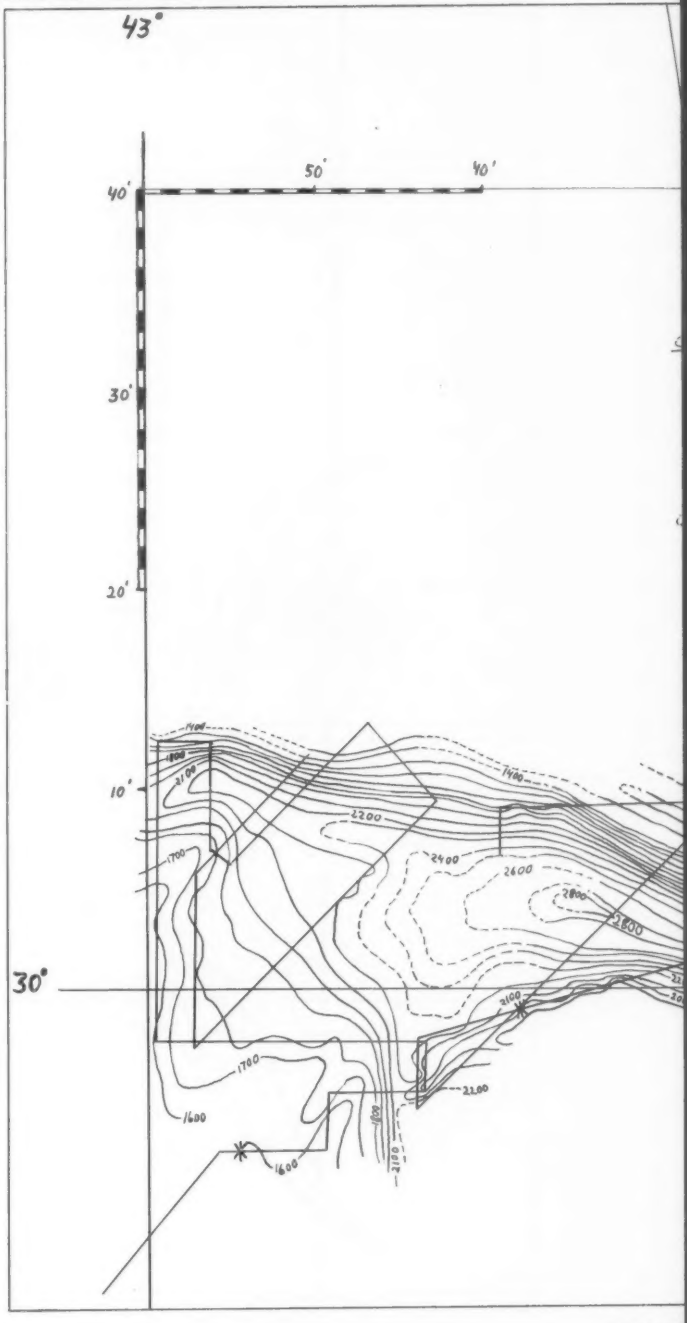

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Main Range of the Mid-Atlantic Ridge. From th
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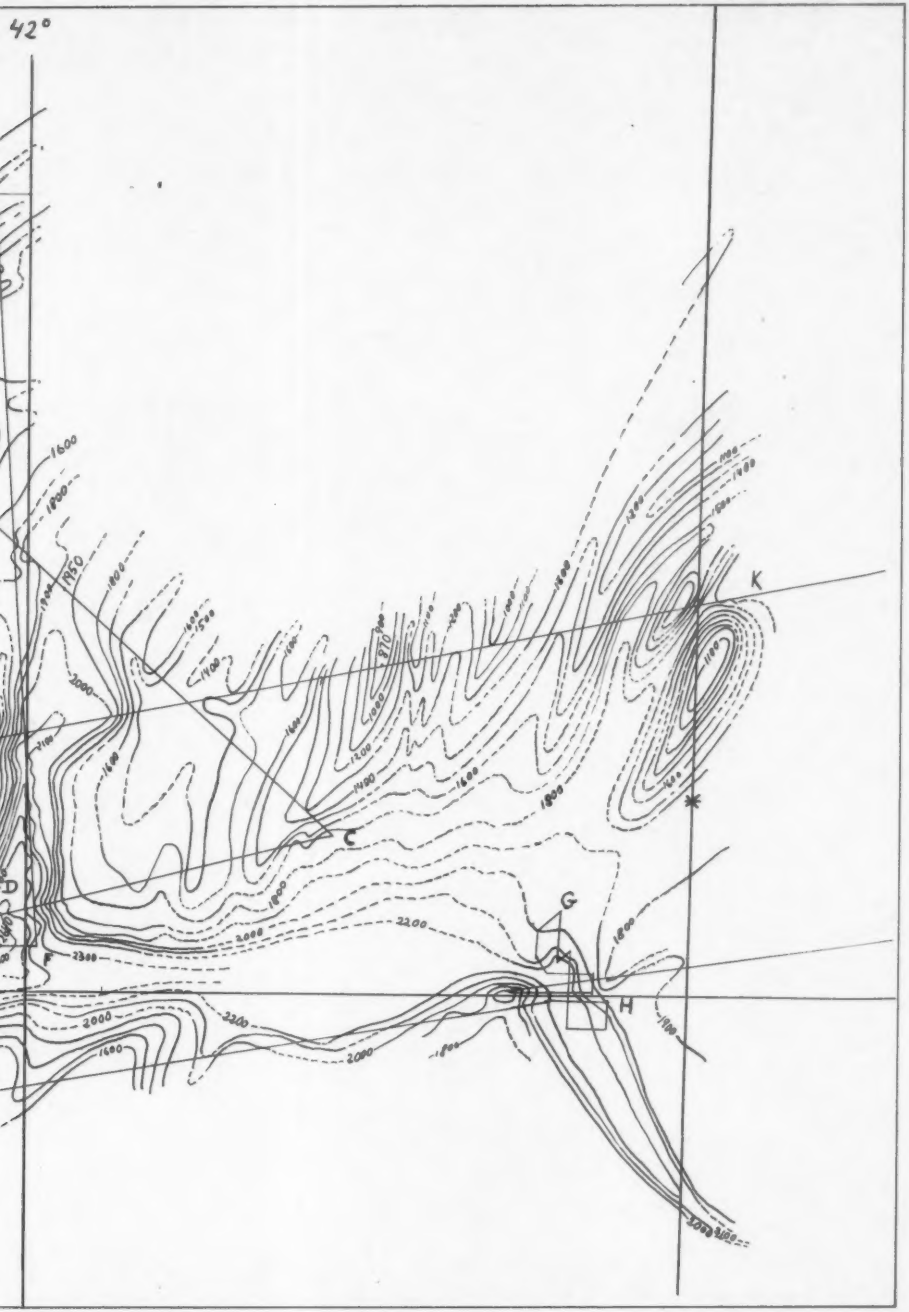
SKETCH MAP OF THE EAST-WEST

ved by the ATLANTIS during the summer of 1947. This trough is also visible on Plates 2 a
nge of the Mid-Atlantic Ridge. From the soundings available, its western and eastern extr
to those on plates 6 c and d. Contour interval, 100 fathoms Asterisks indicate star fixes.



T-WEST TROUGH

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 60, PP. 1541-1596, 11 FIGS. OCTOBER 1949

HAWAIIAN PETROGRAPHIC PROVINCE

By Gordon A. Macdonald

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Abstract

The lavas of the Hawaiian Islands range from mafic picrite-basalts and melilite-nepheline basalts to salic trachytes. Olivine basalt, by far the most abundant type, is regarded as representing the parent magma of the Hawaiian province. Closely associated with the olivine basalts are basalts, and picrite-basalts with many large phenocrysts of olivine. Further differentiation results in eruption of andesine andesite, oligoclase andesite, picrite-basalt with abundant large augite phenocrysts, and more rarely trachyte. Following a long period of quiescence there have been erupted on some islands nepheline basanite, nepheline basalt, melilite-nepheline basalt, "linosaite," and a third type of picrite-basalt. The mineral and chemical composition of the various rock types, and their distribution on the individual islands, are described.

Starting with olivine basalt as the parent magma means of deriving the other rock types are con-

sidered. It is concluded that crystal differentiation has been the principal process, although assimilation of limestone may also have been important. The parts played by gaseous transfer and selective remelting are difficult to evaluate, though both probably operated to some extent.

Comparison of the types of igneous rocks in the Hawaiian province with those recorded from other Pacific islands shows that the rocks throughout the true Pacific Basin are, for the most part, closely similar, and suggests essential uniformity of parent magma and petrogenic processes throughout the Pacific Basin.

Introduction

This paper is the result of petrographic investigations carried on from 1939 to 1949, in connection with the studies of geology and ground-water resources of the Hawaiian Islands by the U. S. Geological Survey in co-operation with the Territory of Hawaii. More than 1500 thin sections have been studied. Other rocks have been examined microscopically by immersion methods, and many others have been studied with a hand lens. The rocks of several islands have been described in detail elsewhere (Macdonald, 1940a; 1940b; 1940c; 1942b; 1946; 1947a; 1947b), and detailed descriptions of the rocks of other islands are in press or preparation. An attempt is here made to sum up the characteristics of the rocks of the entire archipelago, to point out the relationships of the various rock types represented, and to make deductions and suggestions as to their origin.

It is impossible to mention all the persons who have aided the writer's field work in the Hawaiian Islands, but appreciation is expressed to them collectively. Prof. H. S. Palmer of the University of Hawaii kindly supplied the writer with specimens of the rocks collected by him in the Leeward Islands of the Hawaiian Group; Dr. C. K. Wentworth, geologist, and Mr. Frederick Ohrt, manager and chief engineer, of the Honolulu Board of Water Supply, have aided greatly in furnishing laboratory space and equipment for the preparation of thin sections. Dr. Horace Winchell and Doctor Wentworth have made available results of their studies of the Koolau Range Honolulu volcanic series, on Oahu, prior to publication. Dr. H. T. Stearns has aided through discussion of problems and supplying of information throughout the course of the studies. Dr. C. S. Ross has been of inestimable aid in his criticisms of the writer's various manuscripts on Hawaiian petrography.

The three new chemical analyses listed in columns 2, 4, and 5 of Table 6 were made by F. A. Gonyer of Harvard University, under a grant from the Penrose Bequest of The Geological Society of America, through the kindness of the Committee on Hawaiian Petrology, consisting of W. O. Clark, E. S. Larsen, H. S. Palmer, H. A. Powers, H. Winchell, and C. K. Wentworth (chairman).

Previous Work

J. D. Dana, pioneer in many other branches of Pacific geology, published the first (1849) comprehensive descriptions of the rocks of the Hawaiian Islands, although brief descriptions of rocks collected by Goodrich on Kilauea and Mauna Kea volcanoes had been published by Benjamin Silliman (1829; 1831), and Jackson (1846) had published a partial chemical analysis of Pele's Hair from Kilauea. In 1865 Arnold Hague published chemical analyses and brief descriptions of two specimens from Kilauea collected by his brother, J. D. Hague. Brigham (1868) briefly described the rocks and minerals of the Hawaiian Islands and gave partial chemical analyses by Silliman, C. T. Jackson, J. C. Jackson, and J. Peabody.

The early descriptions were based entirely on megascopic features. Although Sorby started studying thin sections of rocks in Europe about 1850, it was not until more than 20 years later that the microscope was applied to the study of Hawaiian lavas. Nepheline basalt from Oahu was described by Wichmann in 1875 and by Rosenbusch in 1877. Glassy basalts from the Hawaiian Islands were briefly described in a preliminary note by Cohen (1876) and more fully by Krukenberg (1877). Cohen (1880) described a suite of specimens collected on the island of Hawaii by William Hillebrand and recognized the presence of andesites. Stelzner (1882) described melilite-nepheline basalt from Oahu. Silvestri (1888) described a suite of specimens collected at Kilauea by P. Tacchini. E. S. Dana (1889) described specimens collected by J. D. Dana from Maui, Oahu, and Kilauea and Mauna Loa on Hawaii; and by E. P. Baker from the summit of Mauna Loa. Merrill (1894) described specimens from Mauna

Kea, collected by E. D. Preston, and (1900) melilite-nepheline basalt from Oahu. Lyons (1896) published chemical analyses and brief descriptions of several lavas and soils from Oahu and Hawaii. Möhle (1902) described specimens collected by H. Schauinsland on Maui, Molokai, Oahu, Kauai, Necker, and Laysan islands. As pointed out by Cross (1915, p. 7) with the exception of those described by Lyons, none of the rocks were collected by the man who described them, and except for J. D. Dana, none of the collectors was a geologist. Consequently the data on the precise source, and the manner of occurrence of the rock body from which the sample was taken, were poor. Dutton (1884) presents only brief summary statements as to the character of the lavas, based on megascopic examinations.

Early in the twentieth century geologic and petrologic studies in the Hawaiian Islands became more systematic. Cross (1904) described the trachyte of Puu Waawaa, on the island of Hawaii. Lindgren (1903) recorded the presence of gabbro on Molokai and Kauai but only very briefly mentioned the character of the lavas of Molokai. Daly (1911) gave careful descriptions and chemical analyses of rocks from Kilauea, Mauna Loa, Mauna Kea, and Hualalai; discussed the petrogenesis; and (1916) summarized the known occurrence of the various types of lavas of Pacific islands. His was the first important contribution to the study of petrogenesis at Pacific volcanoes. Cross (1915) summarized what was known of the petrography of each of the Hawaiian Islands and added new descriptions and chemical analyses of specimens collected by him in 1902, and by C. H. Hitchcock. He discussed the occurrence and relationships of the lavas, their similarities with rocks of other provinces, and the origin of the various rock types. His paper remains of value and is still widely used.

Sidney Powers (1920) published notes on the distribution of the various rock types in the Hawaiian Islands and discussed the origin of the common inclusions of dunite. A series of papers by H. S. Washington (1923a,b,c,d) described lavas from the island of Hawaii collected by him and J. Allan Thomson, presented many new chemical analyses, and discussed the chemical relationships of the lavas. Later papers by Washington and Keyes (1926; 1928) gave descriptions and chemical analyses of rocks from Maui and the Leeward Islands of the Hawaiian group. Stone (1926) carefully described the rocks of Kilauea Volcano. Wentworth (1925) gave brief hand-specimen descriptions of rocks from Lanai; Wentworth and Pegau (1926) described the pyroclastic rocks of Oahu; and Wentworth (1938) published careful descriptions of many specimens of ash and tuff from Hawaii. Palmer (1927; 1930; 1936) described the tuffs of Kaula, Lehua, and Molokini islands. Bowen (1927; 1928) discussed the origin of the olivine-rich picrite-basalts. Hinds (1929; 1930) briefly described the lavas of Kauai and (1925) summarized the knowledge of the nepheline basalts on Kauai and Oahu. Barth (1930) described a feldspar in Hawaiian lavas which he believed to be anemousite, gave (1931b) detailed petrographic descriptions of many of the Hawaiian lavas analyzed by Washington, and (1931a; 1936) discussed the course of crystallization in basalts in general with specific reference to some Hawaiian lavas. Howard Powers (1935) also discussed the crystallization of pyroxenes in Hawaiian lavas and magmatic differentiation in Hawaii.

The U. S. Geological Survey began its systematic studies of the geology of the Hawaiian Islands in 1920, with the investigation of the Kau District on the island of Hawaii. Previous to that time no areal geologic studies in the Hawaiian Islands had been sufficiently detailed and complete to accurately delineate the stratigraphic and structural relationships of the various lava types in the manner needed for a thorough understanding of the petrogenetic processes which have produced them. Since then there has gradually emerged, locally as the result of work by others but largely resulting from the work of the Survey, a fairly complete picture of the structure and stratigraphy of the entire island group. The areal studies by the Survey are now drawing to completion, and there exists for the first time sufficient knowledge of the distribution and relationship of the various rock types to permit profitable speculation as to the petrogenetic processes. The results of the investigations of the Geological Survey on all of the major islands except Kauai have been published in a series of bulletins by

Stearns (1939; 1940a; 1940b; 1947), Stearns and Clark (1930), Stearns and Vaksvik (1935), and Stearns and Macdonald (1942; 1946; 1947). A forthcoming bulletin will describe the geology of Kauai. Discussions of various phases of the petrography by the present writer have appeared separately. (*See* Bibliography.)

Wentworth and Winchell (1947) have described the lavas of the Koolau Volcano on Oahu and Winchell (1947) has studied the lavas of the Honolulu volcanic series. The possible effect of limestone assimilation in the production of Hawaiian lava types has recently been discussed by Daly (1944) and Stearns (1945). Macdonald and Powers (1946) have given detailed petrographic descriptions and new chemical analyses of six lavas of Haleakala Volcano on Maui.

Description of Rock Types

Classification of Rocks

The igneous rocks of the Hawaiian Islands herein are classified chiefly on the basis of the average composition of the modal feldspar. Rocks in which the average feldspar is labradorite or bytownite are termed *basalt* or *gabbro*, according to the granularity. Rocks in which the average feldspar is andesine or oligoclase are termed *andesite*, and those in which alkalic feldspar is dominant are classed as *trachyte*. No Hawaiian lavas contain sufficient quartz to place them in the rhyolite group. Basaltic rocks containing nepheline as well as plagioclase are termed *nepheline basanite*, and nephelinic rocks in which plagioclase is absent are termed *nepheline basalt*. Basanites containing less than 5 per cent modal nepheline or analcite have been termed "*linosaites*" by Winchell (1947). Many basaltic rocks of the Hawaiian Islands contain several per cent of normative nepheline, although no nepheline or other feldspathoid can be detected in the mode. They correspond to the basanitoids of Lacroix. The nepheline is probably occult in the feldspar, but as it is not detectable by microscopic examination it is ignored in the present modal classification. Other basalts contain a little normative quartz, which cannot be detected in the mode, and that likewise is ignored in naming the rock.

Basaltic rocks in which the mafic minerals are so abundant that the total feldspar content is less than 30 per cent are termed *picrite-basalts*. Three distinct types can be recognized, and solely for convenience of reference it is desirable to apply to them distinctive names. One type contains abundant phenocrysts of olivine, but few or none of augite. In other papers the writer has termed this the primitive type of picrite-basalt, because of its association with primitive, or little-differentiated, olivine basalts. It is the oceanite of Lacroix (1927, p. 226). Inasmuch as the olivine-rich picrite-basalts are formed by crystal differentiation, the name "primitive type" is not altogether appropriate and is herein replaced by the designation *oceanite type*. Another type of picrite-basalt contains, in addition to many phenocrysts of olivine, abundant augite phenocrysts. These are the rocks termed ankaramite by Lacroix (1916), and are referred to herein as the *ankaramite type of picrite-basalt*. A third type of picrite-basalt contains moderately abundant but small olivine phenocrysts. Augite phenocrysts are generally absent. The rock is the most mafic of the picrite-basalts. It could be termed melabasalt (Johannsen, 1937, p. 302), but that term applies equally to the oceanite and ankaramite types. It differs from nepheline basalt, tannbuschite, and ankaratrite in the absence of modal nepheline, although normative nepheline is fairly abundant. Cordier's (1868, p. 125; Johannsen, 1937, p. 303) term mimosite appears to have applied to rocks much like those under discussion, which will therefore be called herein the *mimosite type of picrite-basalt*. The groundmass of the oceanite type approximates olivine basalt in composition, indicating that its mafic composition results largely or wholly from the abundant olivine phenocrysts. Both the ankaramite and mimosite types are, however, decidedly more mafic than olivine basalt even when the phenocrysts are disregarded.

Basalts containing less than 5 per cent olivine are termed simply *basalt*, as distinguished from *olivine basalt*, which generally contains 5 to 15 per cent olivine.

The andesites are designated as *andesine andesite* or *oligoclase andesite*, according to the composition of the dominant feldspar. Certain andesites which closely resemble the basalts in habit and in abundance of mafic minerals are termed *basaltic andesite*.

Table 1 shows the average composition of the principal types of lavas in the Hawaiian Archipelago.

Olivine Basalt

Olivine basalt is by far the most abundant rock type in the Hawaiian volcanoes. Most of the olivine basalts are porphyritic, although a relatively few are nonporphyritic. The commonest phenocrysts are olivine, which range up to about 8 mm across but are generally less than 5 mm. The olivine of the phenocrysts has a large negative optic axial angle, probably averaging about 85°, corresponding to a forsterite content of about 80 per cent. Most of the olivine phenocrysts have been partly resorbed, and many altered around the edges and along fractures to iddingsite. In many rocks the iddingsite is enclosed in a shell of fresh olivine which belongs to the period of crystallization of the groundmass, and which tends to restore the euhedral outlines of the partly resorbed crystals. Many olivine basalts contain phenocrysts of plagioclase, from a millimeter to a centimeter, and rarely as much as 5 cm in length. The core of the plagioclase phenocrysts is generally intermediate or sodic bytownite, and in many specimens it is considerably rounded by resorption. The core is surrounded by a narrow, generally strongly zoned rim, in which the composition changes to that of the groundmass feldspar. In a very few flows plagioclase phenocrysts form as much as 30 to 50 per cent of the rock. Pyroxene phenocrysts are much less common than those of olivine and feldspar. Microphenocrysts of pigeonite are found in quite a few olivine basalts, but true phenocrysts of pyroxene are rare. When they do occur they are augite, with an optic axial angle of 55–60°, and generally fairly strong inclined dispersion. Many of them are partly rounded by resorption. Rarely, phenocrysts of olivine, augite, and plagioclase may occur in the same rock. Hypersthene phenocrysts are rare in the lavas of most Hawaiian volcanoes but are fairly common in the olivine basalts of Mauna Loa on Hawaii, and the Koolau Volcano on Oahu, and have been found in the lavas of the Waianae Volcano on Oahu, West Maui, Lanai, Kahoolawe, West Molokai, and Kauai volcanoes. They are typically small, and most are enclosed in a shell of monoclinic pyroxene.

The groundmass of the olivine basalts is generally intergranular or intersertal. Diabasic textures are rare. Hyalopilitic textures are found occasionally, but abundant glass generally is restricted to the thin crusts of pahoehoe flows and cinders and spatter around vents. Interstitial chlorite is rare except in partly altered rocks. Typically, the groundmass consists of plagioclase, monoclinic pyroxene, olivine, and iron ore. The dominant plagioclase is labradorite, generally intermediate or sodic, but rarely calcic. In some rocks there are small anhedral interstitial grains of andesine, and narrow borders of calcic andesine around the labradorite grains. In a few the andesine has a small positive optic angle and is probably potassic (Macdonald, 1942a). In almost all rocks in which its optical properties could be determined, the monoclinic pyroxene is pigeonite. The olivine of the groundmass is generally fresh but sometimes is altered to iddingsite. It contains more iron than does the olivine of the phenocrysts. Determinations of refractive indices of olivine of the phenocrysts and the groundmass in olivine basalts and picrite-basalts of West Maui Volcano show an average content of 82 per cent forsterite in the phenocrysts, as compared with 64 per cent forsterite in the groundmass (Macdonald, 1942b, p. 314). The iron ore includes both magnetite and ilmenite. Both are present in most specimens but magnetite is generally predominant. In a few specimens, however, ilmenite is the predominant opaque mineral, and rarely it appears to be the only one present.

Hypersthene occurs in the groundmass of a few olivine basalts, forming irregular poikilitic grains containing many inclusions of feldspar. The grains are distinctly larger than the other grains of the groundmass, but their relation to the rest of the groundmass does not appear to be that of phenocrysts.

Biotite occurs in many of the olivine basalts, as small irregular interstitial grains in the groundmass and as flakes projecting into vesicles. It is obviously very late magmatic or deuteric. In many rocks apatite forms minute highly acicular crystals enclosed in the feldspars.

TABLE 1.—COMPOSITION OF PRINCIPAL ROCK TYPES OF THE HAWAIIAN PROVINCE

In volume per cent

Name of rock		Olivine basalt	Basalt	Hypersthene-bearing basalt	Picrite-basalt, oceanite type	Picrite-basalt, ankaramite type	Picrite-basalt, mimosite type	Andesine andesite	Oligoclase andesite	Trachyte	Nepheline basanite	Nepheline basalt	Melilite-nepheline basalt
Symbol in Shand's classification*		DVs δ (M'-M")	D(S-V)s δ (M'-M")	D(S-V)s δ (M'-M")	DVs δ M"	DVs δ M"	DVs δ M"	D(S-V)(s-k) γ M'	D(S-V)(s-k) γ M'	D(S-V)(s-k) β (L-M')	DUs δ (M'-M")	DU(s-k) γ (M'-M")	DU(s-k) γ (M'-M")
Megascopic texture		Porphyritic, less commonly nonporphyritic			Porphyritic, phenocrysts large		Porphyritic, phenocrysts small	Porphyritic or nonporphyritic			Porphyritic, rarely nonporphyritic	Porphyritic	
Phenocrysts	Plagioclase	(Ab_{20-50}) 0–25	(Ab_{20-50}) 0–20	(Ab_{20-50}) 0–20	(Ab_{20-45}) 0–5	(Ab_{20-45}) 0–10	— —	(Ab_{20-70}) 0–15	(Ab_{45-80}) 0–5	(Ab_{70-96}) 0–10	(Ab_{25-45}) 0–2	— —	— —
	Nepheline	—	—	—	—	—	—	—	—	—	—	rare	rare
	Augite	0–15	0–3	0–5	0–5	10–30†	—	0–1	—	—	0–5	0–20	0–20
	Hypersthene	—	—	1–10	rare	—	—	—	—	—	—	—	—
	Olivine	0–20	0–3	0–10	20–50†	15–25†	2–20**	0–8	0–5	0–3	0–20	1–20	1–20
	Melilite	—	—	—	—	—	—	—	—	—	—	—	1–35
	Hornblende	—	—	—	—	—	—	—	—?	0–1	—	—	—
	Basaltic hornblende	—	—	—	—	—	—	—	rare	—	—	—	—
	Biotite	—	—	—	—	—	—	—	—	rare	—	—	—
	Riebeckite-like amphibole	—	—	—	—	—	—	rare	0–1	0–1	—	—	—
Microtexture		Generally intergranular or intersertal; average grain size 0.02 to 0.2 mm,							Trachytic, rarely		Intergranular or intersertal;		

	Riebeckite-like amphibole	—	—	—	—	—	—	rare	0–1	0–1	—	—	—

Groundmass	Microtexture	Generally intergranular or intersertal; average grain size 0.02 to 0.2 mm, most commonly about 0.07 mm. Thin chilled crusts and glassy pyroclastic ejecta are hyalopilitic or hyalophitic							Trachytic, rarely intergranular. Glass generally rare. Average grain size usually about 0.04 mm		Intergranular or intersertal; nepheline sometimes poikilitic. Average grain size generally about 0.05 mm		
Composition	Plagioclase	(Ab_{25-50}) 25–45	(Ab_{25-50}) 30–50	(Ab_{25-50}) 27–45	(Ab_{20-45}) 20–30	(Ab_{25-45}) 20–30	(Ab_{20-47}) 10–29	(Ab_{50-70}) 40–55	(Ab_{70-85}) 50–60	(Ab_{90-95}) 65–85	(Ab_{20-50}) 30–50	— —	— —
	Nepheline	—	—	—	—	—	—	—	—	—	1–10	15–45	15–45
	Monoclinic pyroxene	25–45	30–50	30–48	20–35	20–35	45–60	20–40	15–30	5–25	30–40	20–60	20–50
	Melilite	—	—	—	—	—	—	—	—	—	—	—	0–5
	Olivine	1–15	0–3	0–7	1–5	1–5	5–15	1–10	1–8	0–5	0–5	0–5	0–5
	Magnetite and ilmenite	7–15	7–15	8–15	5–15	5–10	7–12	7–20	8–25	5–15	5–8	2–10	2–10
	Apatite	present	present	present	present	present	present	1–2	1–2	1–2	present	present	present
	Rutile	—	rare	—	rare	—	—	—	—	—	—	—	—
	Biotite	—	rare	—	—	—	rare	0–3	0–2	0–2	—	0–2	0–2
	Hornblende	—	—	—	—	—	—	—?	0–1	0–1	—	—	—
	Riebeckite-like amphibole	—	—	—	—	—	—	rare	0–1	—	—	—	—
	Analcite	—	—	—	—	—	0–2	—	—	—	0–8	0–3	0–5

* In assigning the symbols, a small amount of late-crystallized biotite in many of the rocks, especially the andesites, is not regarded as peraluminous. It is probably rich in iron and titania. Likewise the rare hornblende in a few rocks, particularly the oligoclase andesites and trachytes, is probably sodic and is not regarded as metaluminous.

† Phenocrysts large, up to 1 cm long.

** Phenocrysts small, up to 4 mm long.

Picrite-Basalt, Oceanite Type

Picrite-basalts of the oceanite type differ from the olivine basalts only in the presence of numerous large phenocrysts of olivine. The latter generally reach lengths of 5–8 mm but may be as much as 1 cm or more in length. As in the olivine basalts, the olivine phenocrysts have a large negative optic angle, corresponding to a forsterite content of about 80 per cent. This agrees with the composition Fo_{81} determined by chemical analysis for the olivine phenocrysts in the picrite-basalt of the 1840 lava flow of Kilauea (Aurousseau and Merwin, 1928). The olivine phenocrysts are generally partly resorbed, and many are partly altered to iddingsite. They commonly constitute 25 to 30 per cent of the rock and in rare instances reach as much as 50 per cent. Through a decrease in the abundance of olivine phenocrysts, the rocks grade into the olivine basalts. Augite and plagioclase phenocrysts occur in some flows but are decidedly subordinate to those of olivine. Rare hypersthene phenocrysts were found in one rock from Kauai and one from West Maui. The composition and textures of the groundmass are identical to those of the olivine basalts. Brown biotite has been found as a late-magmatic constituent in two picrite-basalts of the oceanite type on Kauai.

Picrite-Basalt, Ankaramite Type

The ankaramite type of picrite-basalt differs from olivine basalt principally in the abundance of augite and olivine phenocrysts; but even disregarding the phenocrysts the groundmass is more mafic than are the olivine basalts (Table 9). Phenocrysts of olivine and augite generally constitute 25 to 35 per cent of the rock but may form 50 per cent, or a little more. In most specimens the olivine and augite phenocrysts are approximately equal in abundance, but in some either one or the other may predominate. Augite phenocrysts commonly reach a length of 7–10 mm, and rarely as much as 2 cm. The olivine phenocrysts are generally a little smaller. Both minerals are commonly partly resorbed, and the olivine may be partly altered to iddingsite. The olivine has a large negative optic angle. The augite has an optic angle of 55–60°, moderate to strong inclined dispersion, and a β refractive index of about 1.700. Typical augite phenocrysts from a lava flow near the top of Haleakala Volcano have the following composition (Washington and Merwin, 1922):

	per cent		per cent
SiO_2	47.70	K_2O	0.03
Al_2O_3	6.82	TiO_2	1.89
Fe_2O_3	3.36	Cr_2O_3	0.23
FeO	4.43	MnO	0.16
MgO	13.34	H_2O+	0.13
CaO	21.35		
Na_2O	0.65	Total	100.11

Feldspar phenocrysts, resembling those in the olivine basalts, may be present. In some rocks they are abundant, resulting in a high total feldspar content and gradations into the olivine basalts. The ankaramite type of picrite-basalt grades into olivine basalt through the decrease in abundance of mafic phenocrysts. The groundmass textures and composition are like those of the olivine basalts. In all specimens in which its optical properties can be determined, the groundmass pyroxene is pigeonite, and narrow rims of pigeonite surround many of the phenocrysts of augite.

Picrite-Basalt, Mimosite Type

All specimens of the mimosite type of picrite-basalt contain phenocrysts of olivine, but the phenocrysts are small, and generally only moderately abundant. In most rocks they are less than 1.5 mm long, and no phenocrysts seen exceeded a length of 4 mm. In hand specimen they are dark and resemble those of the nepheline basalts. Generally they are partly resorbed. Many are altered around the edges to iddingsite, and in some rocks the iddingsite is surrounded by a thin envelope of fresh olivine. Augite phenocrysts are very rare; none have been found exceeding 1 mm in length. They are enclosed in thin shells of pigeonite. Microphenocrysts of pigeonite are somewhat more common but are probably not of intratelluric origin.

The groundmass consists largely of pigeonite, with less abundant plagioclase, olivine, and iron ore. Apatite occurs as inclusions in the feldspar. Interstitial glass is present in some rocks, and a few contain small interstitial flakes of brown biotite. The pigeonite is colorless to pale brown in thin section; but in some in-

stances, in the vicinity of vesicles or in small pegmatitoid patches, it exhibits purplish-brown titanian borders. The plagioclase is intermediate or calcic labradorite (β = 1.560 to 1.567). In some rocks it forms anhedral interstitial grains between subhedral grains of pigeonite, olivine, and magnetite; but in others it forms large anhedral grains which enclose the mafic constituents. In most rocks feldspar constitutes about 20–25 per cent, but in some it forms as little as 10 per cent. In others, an increase in the proportion of feldspar results in a gradation into the olivine basalts. Rarely, a few grains of analcime are present. A typical specimen, collected on the eastern side of Lawai Bay, on Kauai, consists of labradorite, 23 per cent; pigeonite, 45 per cent; magnetite, 7 per cent; olivine phenocrysts, 7 per cent; and groundmass olivine, 18 per cent.

The mimosite type of picrite-basalt has been found only on Kauai.

Basalt

The basalts differ from the olivine basalts only in their smaller percentage of olivine, which ranges from the arbitrary upper limit of 5 per cent to less than 1 per cent. Nonporphyritic rocks are more abundant than among the olivine basalts, but most rocks contain a few small phenocrysts of olivine or plagioclase. The olivine phenocrysts are generally partly resorbed. The groundmass is similar in composition and texture to that of the olivine basalts, except that olivine is scarce or lacking. Wherever its nature could be determined, the groundmass pyroxene is pigeonite. The average plagioclase composition is about the same as in the olivine basalts. Basalts free of olivine are rare, but have been found among the lavas of several Hawaiian volcanoes. Hypersthene is rare in the olivine-poor and olivine-free basalts; its occurrence is the same as in the olivine basalts. Biotite also occurs, as in the olivine basalts.

Andesine Andesite

The typical andesine andesites are gradational on the one side with the basalts, from which they differ in the more sodic nature of their average feldspar and generally in the smaller percentage of mafic minerals, and on the other side with the oligoclase andesites. Many of them show a somewhat trachytic texture, and in many a well-developed platy jointing parallels the flow planes (and hence in general parallels the top and base of the flow). The platy joints exhibit an almost schistose sheen owing to reflection from the crystals and cleavage faces of innumerable small tabular crystals of feldspar. Similar platy jointing is characteristic of the oligoclase andesites.

The andesine andesites are typically nonporphyritic, but some flows contain small phenocrysts of feldspar and less commonly of olivine. Augite phenocrysts have not been found, but microphenocrysts of pigeonite are present in a very few rocks. The olivine phenocrysts are generally partly resorbed, and in some rocks they are partly altered to iddingsite. The feldspar phenocrysts are typically shorter and more blocky in habit than those of the basalts. They contain cores, which range from intermediate bytownite to labradorite, enclosed in thin shells of intermediate or sodic andesine. In many specimens the cores are rounded by resorption, and the transition from the core to the more sodic rim is generally sharp. Rarely, the andesine andesites contain small, partly resorbed phenocrysts of a riebeckitelike mineral, which will be described in connection with the oligoclase andesites.

The groundmass of the andesine andesites is typically intergranular, or more rarely intersertal, with moderately to well-developed fluidal structure. The feldspar of the groundmass is intermediate or sodic andesine. In many rocks the late feldspar shows a small positive optic angle and is probably potassic. The pyroxene, wherever its properties can be determined, is pigeonite. There is generally a little olivine in the groundmass, although in a few specimens it is absent. Iron ores are in general a little more abundant than in the more mafic lavas. Magnetite is distinctly more abundant than ilmenite and in some rocks appears to be the only opaque mineral. Apatite is present as minute acicular crystals, generally enclosed in the feldspar. Biotite is widespread and, as in the basalts, is clearly of late magmatic or deuteric origin. Hornblende is very rare.

Many of the andesine andesites of the

Hawaiian Islands resemble the basalts in habit and in the abundance of mafic minerals, differing from them only in the more sodic nature of their average feldspar. In hand specimen many of them are indistinguishable from the olivine-poor basalts. They have been termed basaltic andesites.

Oligoclase Andesite

Many flows of oligoclase andesite contain phenocrysts of feldspar, some of them as much as 8 mm long, but most flows are nonporphyritic. The feldspar phenocrysts consist of a core with composition ranging in different rocks from intermediate labradorite to intermediate oligoclase, surrounded by a thin shell having the composition of the groundmass feldspar. Small phenocrysts of olivine occur in a few flows. Augite phenocrysts have been found in only one specimen, from Lalakea Gulch on Kohala Volcano on Hawaii. Partly resorbed phenocrysts of basaltic hornblende are found in several flows.

Many specimens of oligoclase andesite contain small phenocrysts of a mineral resembling riebeckite, but less strongly colored. The mineral is distinctly pleochroic, with X = grayish-brown or bluish-gray, and Z = pale yellowish-brown to yellow. It has good prismatic cleavage, nearly parallel extinction, negative elongation, moderate refringence, very low birefringence (.004±), and 2V (+) about 75 . The mineral has also been observed in the groundmass of a few specimens.

The texture of the oligoclase andesites is trachytic. The groundmass consists predominantly of feldspar, with smaller amounts of colorless monoclinic pyroxene, iron ore, and a little olivine. The groundmass of the oligoclase andesites of Kohala Volcano is on the average a little more basic than that of the oligoclase andesites of West Maui Volcano; in the lavas of Kohala the feldspar is generally calcic oligoclase, whereas in the lavas of West Maui it is generally sodic oligoclase. Moreover, in the Kohala rocks the proportion of mafic minerals is generally a little greater than in those of West Maui. In these respects, and also in the presence of a little aegirine-augite, some of the oligoclase andesites of West Maui are transitional into the trachytes. In many specimens the granularity is too fine to permit determination of the optical properties of the pyroxene, but in all except one in which they could be determined the mineral is pigeonite. Biotite is common and forms irregular interstitial grains in the groundmass or projects into vesicles. Hornblende is present in the groundmass of a few rocks. Apatite forms acicular grains, generally minute and enclosed in the feldspar, but in a few rocks it exceeds 1 mm in length. The iron ore includes both magnetite and ilmenite, with magnetite greatly predominant.

Trachyte

The trachytes are gradational into the oligoclase andesites and differ from them principally in the more sodic nature of the feldspar and the smaller proportion of mafic minerals. The trachytes are light-gray dense rocks, generally with moderately well developed flow banding, and platy jointing parallel to the flow planes. The joint surfaces show the same almost schistose sheen noted in the andesites. Nearly all specimens contain blocky phenocrysts of feldspar 2–7 mm long. The feldspar phenocrysts are calcic to intermediate oligoclase and are surrounded by a narrow rim in which the composition changes rapidly to albite. A few flows of trachyte contain phenocrysts of hornblende. Phenocrysts of basaltic hornblende and biotite occur in the trachyte of Mauna Kuwale, in the Waianae Volcano on Oahu. A few hypersthene phenocrysts, partly replaced by hornblende, also are present in the Mauna Kuwale trachyte.

The texture of the trachytes is trachytic. The groundmass consists largely of subhedral to anhedral grains of albite. Ferromagnesian minerals, which generally constitute less than 15 per cent of the rock, include aegirine-augite, hornblende, biotite, magnetite, the riebeckite-like amphibole, and rarely acmite. The groundmass of the trachyte of Mauna Kuwale is exceedingly fine-grained and consists of tiny microlites of oligoclase (?) in a matrix of alkali feldspar and glass. The trachytes of West Maui Volcano and Puu Anahulu, on Hawaii, contain a little normative nepheline. They are essentially saturated in silica. The trachyte of Mauna Kuwale, in contrast, contains nearly 20 per cent of normative quartz and distinctly less alkalies than the trachytes of West Maui and Hawaii (Macdonald, 1940a, p. 84).

Nepheline Basanite

The nepheline basanites resemble the olivine basalts and picrite-basalts except that the place of some of the feldspar is taken by nepheline. Most contain phenocrysts of olivine, which in a few constitute 15 or 20 per cent of the rock. Many also contain augite phenocrysts, but these are seldom abundant. The groundmass is intergranular or intersertal, consisting of labradorite, pigeonite, and iron ore, with or without interstitial glass. Olivine is present in the groundmass of some flows. A small amount of analcime is found in a few specimens. Feldspar is several times as abundant as nepheline. Gradations into the nepheline basalts, through decrease in feldspar and increase in feldspathoid, are not known.

Gradations into the olivine basalts and the picrite-basalts of mimosite type have been found. On Oahu olivine basalts containing very small amounts of nepheline or analcime have been termed "linosaite" by Winchell (1947), and similar rocks are present on Kauai. Winchell has classified several flows on Oahu as "linosaite," on the basis of their association with nepheline basalts and basanites, even though the presence of feldspathoids cannot be demonstrated. However, definite examples of gradational rocks are rare.

Nepheline Basalt

Phenocrysts of olivine are present in all specimens of nepheline basalt, ranging from a few scattered individuals to 20 per cent of the rock. They are generally less than 2 mm long but occasionally are more than 6 mm. The olivine phenocrysts of the nepheline basalts are generally darker than those of the typical olivine basalts, and somewhat richer in fayalite, although the iron content is not as high as in the olivines of the oligoclase andesites (Macdonald, 1942b, p. 314). In some rocks the olivine phenocrysts are fresh, but in many they are partly altered to iddingsite, and in some the iddingsite is surrounded by a thin jacket of fresh olivine. Rarely the phenocrysts are sharply euhedral, but more commonly they are partly resorbed. In some rocks the resorbed phenocrysts are surrounded by a thin layer of finely granular exsolved iron ore, and in a few from Kauai they show narrow coronas of monoclinic pyroxene admixed with iron ore.

Phenocrysts of augite are common in the nepheline basalts of Oahu, but rare on Kauai. They are purplish brown, with strong dispersion, and hence titanian. Zoning and hour-glass structure are common. The outer bands have a smaller optic axial angle and extinction angle than the core, representing an increase in iron content. Microphenocrysts of pigeonite are present in a few nepheline basalts from Kauai.

Nepheline phenocrysts are very rare, but have been found at three localities on Kauai, at one of which they attain lengths of 1.5 mm. They are much rounded by resorption.

The groundmass of the nepheline basalts consists essentially of nepheline, pigeonite, and iron ore. Melilite may be present, resulting in melilite-nepheline basalts. Three general types of groundmass texture can be recognized. The most common is composed of subhedral grains of pigeonite, iron ore, and in some specimens olivine, between which lie small anhedral grains of nepheline. Less widespread, but still common, are rocks in which large anhedral grains of nepheline enclose subhedral grains of pigeonite, iron ore, and olivine. Least common is a texture in which the abundant small grains of nepheline are subhedral, showing nearly square outlines in cross section, with the grains of other minerals partly enclosed in them and partly lying between them.

In some rocks the pigeonite is very pale brown or colorless, but in many it is purplish, with strong dispersion, and is probably titanian. Rarely, it grades outward into borders of aegirine-augite. Olivine is present in the groundmass of some rocks, but lacking in others. The iron ore grains are largely magnetite, but probably are titanian. Ilmenite is present in some specimens. Minute highly acicular grains of apatite are common, generally enclosed in the nepheline. Irregular flakes of reddish-brown or purplish biotite are common in the nepheline basalts of Kauai, but rare on Oahu. They are very late magmatic or deuteric and occupy interstices between the other grains or project into vesicles. Small grains of brown hornblende observed in a few specimens likewise appear to be of late crystallization. Interstitial chlorite or serpentine is present in a few rocks, and some contain a little interstitial glass. Brown isotropic grains of perovskite are present in a few

nepheline basalts. Calcite and zeolite occur quite commonly in vesicles. Analcime is recognized in some rocks. It is generally primary, but in some rocks is an alternation product of nepheline.

Melilite is present in small amounts in many of the nepheline basalts, and in some forms as much as 20 to 30 per cent of the rock. Where it is abundant, it forms small phenocrysts, but in many lavas it is restricted to a few small grains in the groundmass. Lavas containing it are termed melilite-nepheline basalts. In other respects they are much like the nepheline basalts. Some of the melilite phenocrysts show zoning with the outer bands a little richer in akermanite than the core. Winchell (1947, p. 24) determined the refractive indices of melilite phenocrysts in a lava from Oahu to be $\omega = 1.661$ and $\epsilon = 1.655$, indicating about 75 per cent gehlenite and 25 per cent akermanite. Melilite from the groundmass of a lava collected 1.7 miles southwest of Koloa, on Kauai, is optically negative, with $\omega = 1.659$, indicating a content of about 67 per cent gehlenite and 33 per cent akermanite. The birefringence is low, but ultra-blue absorption colors are rare. Peg structure is very rare. In some lavas the melilite has been altered to pale honey-yellow or brown isotropic material.

On both Oahu and Kauai, the nepheline basalts contain pegmatitoid veinlets and irregular segregations. These are coarser-grained than the enclosing rock but at many localities grade into it and are obviously related to it genetically. They represent the last-crystallized portion of the nepheline basalt magma, rich in volatiles, accumulated in irregular pockets or pressed out into fractures. Besides the minerals typical of the parent rock, they contain others of deuteric and hydrothermal origin. The pegmatitoid veinlets have been studied in detail by Dunham (1933; 1935) and Winchell (1947, p. 26). Winchell lists the following paragenetic association:

Pyrogenic	Chiefly magmatic: olivine, augite, nepheline, melilite, magnetite.
	Chiefly deuteric: perovskite, apatite, hastingsite, analcime.
Secondary	Metacolloid: allophane.
	Zeolitic: chabazite, phillipsite, hydronepheline, quartz, calcite.

Many of the small pegmatoid patches are less complex mineralogically than the veinlets studied by Dunham and Winchell. The simplest contain only those minerals found in the enclosing rock and are distinguished only by their coarser granularity. Other more complex pegmatitoids contain deuteric as well as magmatic minerals, but lack the zeolites and metacolloids. As an example may be cited small roughly spherical and lenticular patches in a nepheline basalt exposed 0.42 mile N. 22° W. of benchmark 436 at Lawai, Kauai. They are much coarser than the enclosing rock and contain the same magmatic minerals—augite, olivine, nepheline, and magnetite. Many of the augite grains grade into borders of aegirine-augite. In addition there occur a little perovskite and biotite, small grains of aegirite, and highly acicular grains of an amphibole, pleochroic from yellowish brown to deep reddish brown, with the greatest absorption normal to the direction of elongation, and low birefringence (probably kataphorite). A few of the patches also contain zeolite and calcite.

A rare variation of the nepheline basalts is the ultramafic type termed "ankaratrite" by Lacroix (1916). In this type the mafic minerals are very abundant, and nepheline constitutes only a small percentage. There is a complete gradation from the ordinary nepheline basalts, containing 30–45 per cent nepheline, to rocks containing as little as 15 or 20 per cent. The "ankaratrites" closely resemble the picrite-basalts of the mimosite type, except that the light-colored constituent is nepheline instead of feldspar. They constitute a gradation from the typical nepheline basalts to the mimosite type of picrite-basalt. Like the latter, they have been found only on Kauai.

Gabbro

Small stocks of gabbro have been exposed by erosion in several of the older Hawaiian volcanoes, and at others fragments of gabbro carried up from depth are enclosed in lava flows or occur among the pyroclastic ejecta. The gabbros are the normal plutonic or hypabyssal equivalents of the olivine basalts and picrite-basalts. Their texture is granitoid. Some are dense, but many have an open miarolytic structure (Macdonald, 1942a, Pl. 43c). They

consist of plagioclase (bytownite to labradorite), monoclinic pyroxene, iron ore, and usually olivine. The pyroxene is generally augite, but in some, particularly those with fairly fine granularity, it is pigeonite. Some contain interstitial oligoclase or andesine which has a small positive optic angle and is probably potassic. In others the interstitial feldspar is antiperthite, consisting of an intergrowth of soda-lime feldspar with potash feldspar. An ejected block from a cone on the western side of Mauna Kea is nearly devoid of pyroxene but contains abundant olivine and may be classed as a troctolite.

A small laccolith of picritic gabbro porphyry is exposed in the wall of Kilauea Caldera at Uwekahuna Bluff (Daly, 1911, p. 291). The rock of the Uwekahuna laccolith, and ejected blocks of similar rock at the summit of Hualalai Volcano, are finer-grained than the typical gabbros and contain numerous phenocrysts of olivine.

Ultrabasic Rocks

Inclusions of dunite are common in Hawaiian lavas. They are dense oil-green to brownish-green rocks, with a sugar-granular texture. The average grain size commonly is about 1.5 mm, although in some specimens it is as much as 5 mm. The texture of some of the coarser-grained specimens has a faintly clastic appearance, as though the olivine grains had been partly crushed, leaving partly rounded residuals in a matrix of finer grains. The dunite consists almost entirely of olivine, with a few subhedral grains of a metallic spinel, probably magnetite, and in some specimens a few grains of augite.

Blocks of augite peridotite (wehrlite), consisting of about 75 per cent olivine and 25 per cent augite, occur on the summit cinder cone of Mauna Kea. By an increase in feldspar content, these rocks grade through picrites into the gabbros. Inclusions of augite peridotite in the lava flow of 1801 from Hualalai show gradations into the dunites. A dunite inclusion from West Maui Volcano is transected by a band of augitite, 2.5 cm wide, the contacts between the dunite and the augitite being gradational. Many of the augite peridotites exhibit similar banding on a slightly smaller scale, layers of nearly pure olivine alternating with thinner layers rich in augite. The banding resembles that found in many norites and peridotites and is probably the result of simultaneous crystallization and rhythmic differential settling of the two minerals (Coats, 1936; Hess, 1938).

Blocks of augite-hypersthene peridotite (lherzolite) enclosed in an oligoclase andesite of Kohala Volcano in the sea cliff near Kukuihaele contain hypersthene as well as augite and olivine. The hypersthene has $\beta = 1.698$, indicating a content of about 76 per cent enstatite. Bombs of garnetiferous augite-hypersthene peridotite from Aliamanu Crater on Oahu, termed "olivine eclogite" by Winchell (1947), contain in order of abundance augite, olivine, orthorhombic pyroxene, garnet, magnetite, and biotite (Wentworth and Pegau, 1926). According to Winchell (1947), the garnet is of two varieties. One consists of about 40 per cent almandite and 60 per cent pyrope, probably with some grossularite; the other consists of about 80 per cent almandite and 20 per cent pyrope, probably with some andradite.

Distribution and Structural Relationships of Rocks

Leeward Islands

The Leeward Islands of the Hawaiian Archipelago are the several small islands which lie along a west-northwestward trending line west of Kauai and Niihau. From Kure (Ocean) Island at the west (Fig. 1), they include Midway Island, Pearl and Hermes Reef, Lisianski Island, Laysan Island, Gardner Island, French Frigate Shoal, Necker Island, Nihoa (Bird) Island, Kaula Island, and several other reefs and shoals. On most of them only coralline limestone and calcareous sands are exposed, but the reefs are presumably built upon a volcanic base. Volcanic rocks have been found only on Laysan Island, Gardner Island, French Frigate Shoal, Necker Island, Nihoa Island, and Kaula Island.

Comparatively little is known regarding the petrography of the volcanic islands of the Leeward Group. Möhle (1902, p. 90–93) described a few specimens from Laysan and Necker, and Powers (1920, p. 257) described single specimens from Necker and Nihoa. Washington and Keyes (1926, p. 340–352) described samples from Necker, French Frigate Shoal, and Nihoa,

collected by H. S. Palmer, and a single specimen from Gardner, collected by S. C. Ball. Several of the specimens were analyzed chemically. The geology of French Frigate Shoal

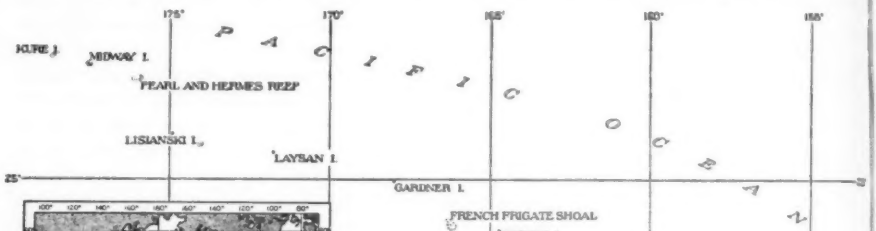


FIGURE 1.—MAP OF THE HAWAIIAN ARCHIPELAGO

Showing the location of the major islands. Small inset map shows the position of the Hawaiian Islands in the Pacific Ocean.

and Kaula, Nihoa, Necker, and Gardner islands has been discussed by Palmer (1927; 1936), who describes several samples from each island in addition to those described by Washington and Keyes. Through the kindness of Professor Palmer, the writer has re-examined specimens of most of the rocks analyzed by Washington and Keyes.

Kaula Island is a crescentic fragment of a tuff cone, composed of partly palagonitized basaltic tuff, at the summit of a broad, wave-truncated, submarine shield volcano. In the tuff Palmer (1927; 1936) identified fragments of olivine, augite, magnetite, and glass. Because of the abundant glass it is not certain, lacking a chemical analysis, whether the rock is a basalt or a basanite. Blocks of reef limestone, calcareous sandstone, nonporphyritic basalt, olivine basalt containing olivine phenocrysts, and bombs with cores of dunite are enclosed in the tuff.

Lehua Island, just north of Niihau, is a cone of palagonitized basaltic tuff, much like Kaula Island. The tuff contains blocks of both porphyritic and nonporphyritic olivine basalt, and the later layers contain blocks of reef limestone (Palmer, 1936).

The rocks of Nihoa appear to be predominantly olivine basalts. Olivine-poor and olivine-free basalts, and at least one flow of picrite-basalt of the oceanite type also are present. The rocks of Necker Island also appear to be largely olivine basalts, but one of the specimens analyzed by Washington and Keyes is a basaltic andesite, and several flows are picrite-basalt of the ankaramite type. The chemical analysis of a 3-foot dike (Washington and Keyes, 1926, p. 347) indicates the presence of 19.3 per cent normative nepheline, although no modal nepheline is present. The rock was classed by Barth (1931b, p. 515) with the phonolites, but its silica content of 47.89 per cent is much lower than that of typical phonolite. Chemically the rock resembles the essexites and tahitites of Tahiti, except that it contains less potash. It lacks the hauyne characteristic of the tahitites. Isotropic material in the groundmass is faintly blue when stained by the method described by Shand (1939), and is probably analcime. It appears best to class the rock with the basanites.

The only volcanic rocks exposed on French Frigate Shoal form La Perouse Rock. The two specimens collected by Palmer are olivine basalts. The lava flows and dikes of Gardner

Island also are basalts. Some contain small phenocrysts, but others are nonporphyritic, and the specimen analyzed by Washington and Keyes (1926, p. 349–350) was reported to be olivine-free. Blocks of gabbro are enclosed in a bed of tuff.

Specimens of olivine basalt, containing large phenocrysts of olivine, were collected by Schauinsland from three large blocks found in 1–2 meters of water on the reef of Laysan Island. Other specimens of basalt were collected from beach boulders (Möhle, 1902, p. 92–93).

Thus most of the rocks of the Leeward Islands are olivine basalt. Evidence of extensive differentiation is found only on Necker Island, where, in addition to the olivine basalts, basaltic andesite and picrite-basalts of the ankaramite type and at least one dike of basanite are present. The picrite-basalt of oceanite type recorded from Nihoa represents only the slight degree of differentiation which normally occurs among the early lavas of the basaltic shield volcanoes such as Kilauea.

Niihau

The east-central part of Niihau Island is a remnant of a basaltic shield volcano, built probably during the Tertiary period. Its rocks, named the Paniau volcanic series (Stearns, 1947, p. 14), consist of thin flows, predominantly of olivine basalt, but including a smaller proportion of basalt and picrite-basalt of the oceanite type. A little basaltic andesite is probably present (Macdonald, 1947b). Innumerable northeast-trending dikes, of the same types as the flows, cut the lavas.

The cessation of the principal period of volcanism was followed by a long period of erosion and weathering, during which stream valleys were formed and waves cut a broad platform along the western side of the island. Much of the eastern portion of the volcano disappeared, owing to wave action, downfaulting, or both. Later, probably during the Pleistocene epoch, lavas erupted on the wave-cut platform built the low plain which now borders the higher remnant of the shield volcano on the north, west, and south (Fig. 2). The rocks of the late eruptive series have been named (Stearns, 1947, p. 14) the Kiekie volcanic series. Five of the vents of the Kiekie volcanic series are aligned on a nearly north-south fissure crossing the western side of the island. Most of the lavas are olivine basalt, but one is transitional from olivine basalt to picrite-basalt. Melilite-nepheline basalt has been reported from one locality (Hinds, 1925, p. 534), but its occurrence has not been confirmed. The northernmost of the Kiekie vents now exposed above sea level built the tuff cone of Lehua Island.

Kauai

The main bulk of the island of Kauai consists of a deeply dissected shield volcano (Fig. 3). A large caldera was once present, but appears to have been asymmetrical, the boundary cliffs on the south and west sides being much higher than those on the north. The cliffs on the south and west sides confined the lava flows within the caldera long after the flows had buried and overflowed the boundary cliffs at its northern edge. In places along the southern and western borders of the caldera the contact of the thick horizontally bedded intracaldera flows against the thinner outward-dipping extracaldera flows is strikingly exposed (Stearns, 1946, p. 85); but, in general, it has not proved possible to map a caldera-filling member. On the northern edge the lavas appear to have overtopped the boundary cliffs of the caldera and poured northward over the outer slope of the volcano at a level considerably lower than that now exposed by erosion.

The name Waimea Canyon volcanic series is herein proposed for the rocks of the major volcanic mountain of Kauai, after the type locality in the walls of Waimea Canyon, where both intracaldera and extracaldera flows are excellently exposed in sections as much as 2600 feet thick, and their unconformable relationship to the lavas of the later Koloa volcanic series is clearly revealed.[1]

On the southeastern flank of the major volcano lies a smaller shield volcano with a filled caldera, very deeply dissected. The rocks of this smaller shield appear to interfinger with those of the major shield.

The lavas of both these shields are principally

[1] The rocks herein termed the Waimea Canyon volcanic series were called by Stearns (1946, p. 85) the Waimea volcanic series. However, the name Waimea is preoccupied, and it is therefore necessary to assign a new name to the group.

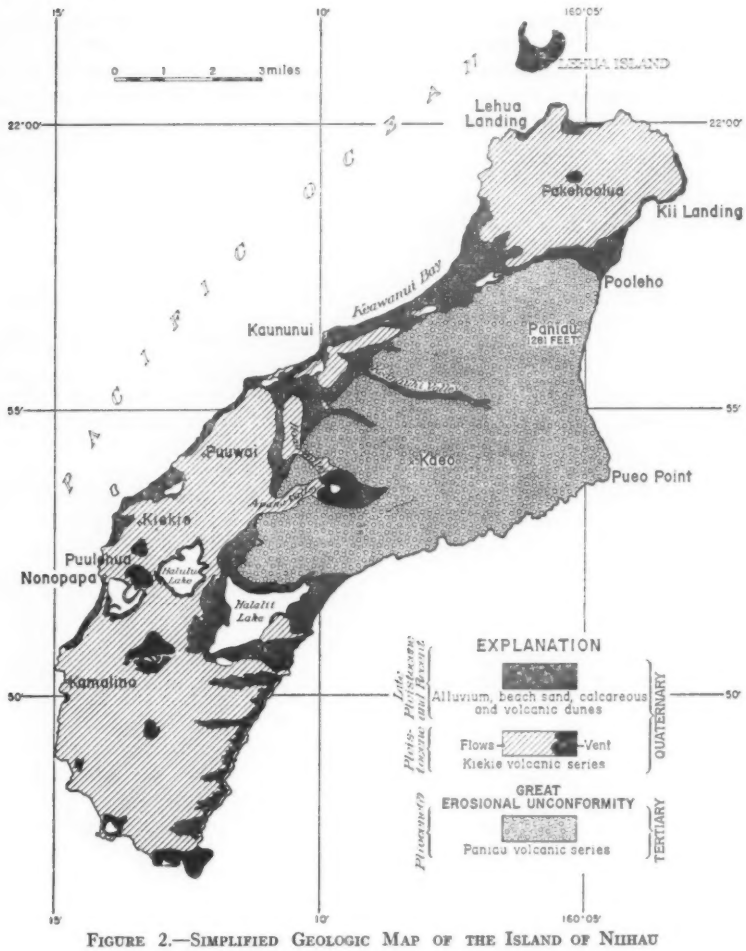


FIGURE 2.—SIMPLIFIED GEOLOGIC MAP OF THE ISLAND OF NIIHAU
Slightly modified after H. T. Stearns (1946, Fig. 24).

olivine basalt, but basalt and picrite-basalt of the oceanite type also are present throughout the section. A few of the basalts and olivine basalts, and rarely the picrite-basalts, contain a small amount of hypersthene. A few flows of andesine andesite have been found in the uppermost part of the section. Small gabbro stocks cut the lavas. Most of the gabbros are normal in composition, but one type contains titanian augite, aegirite, anorthoclase, and orthoclase, and has been described by Cross (1915, p. 14) under the name "kauaiite." It probably is related genetically to the lavas of the Koloa volcanic series.

After a long period of volcanic quiescence and deep erosion, volcanism was renewed with the extrusion of lava flows and building of small shield volcanoes and cinder cones, and at least one tuff cone. This period of eruption was called by Hinds (1930, p. 57) the Koloa volcanic episode, and its rocks are termed (Stearns, 1946, p. 85–89) the Koloa volcanic series. The Koloa eruptions continued over a long period, inundating much of the southern, eastern and

FIGURE 3.—SIMPLIFIED GEOLOGIC MAP OF THE ISLAND OF KAUAI

Geology by H. T. Stearns, G. A. Macdonald, D. A. Davis, and D. C. Cox, 1939–1949.

northeastern parts of the island (Fig. 3). In some places long periods of weathering and erosion separated successive flows of the Koloa volcanic series (Clark, 1935), but there does not appear to be any general island-wide division into two members separated by a period of volcanic quiescence. Probably volcanism was active locally, for short periods, while weathering and erosion continued in the rest of the area. The flood of lavas of the Koloa volcanic series which inundated the lowlands of Kauai after the long post-Waimea Canyon period of erosion greatly exceeds the volume of the similar late eruptions of the Honolulu volcanic series on Oahu.

The lavas of the Koloa volcanic series include olivine basalt, "linosaite," picrite-basalt of the mimosite type, "ankaratrite," nepheline basanite, nepheline basalt, melilite-nepheline basalt, and at least one flow of analcime basanite. Many of the nepheline and melilite-nepheline basalts contain moderately abundant biotite. Dunite inclusions are numerous in some of the flows.

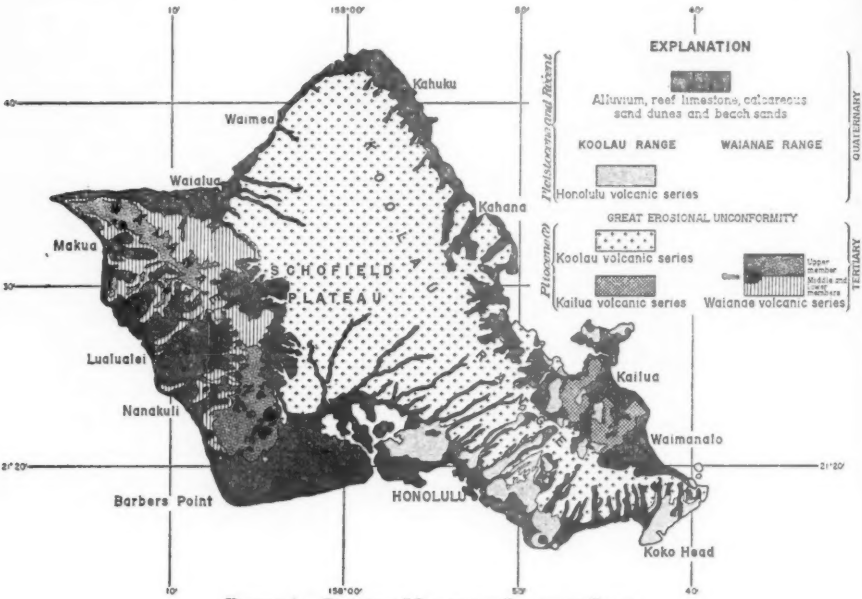


FIGURE 4.—GEOLOGIC MAP OF THE ISLAND OF OAHU
Simplified after Stearns, 1939, Pl. 2 and Macdonald 1940a, Pl 6.

Oahu

Waianae Volcano.—Waianae Volcano, which forms the western part of Oahu (Fig. 4), is the older of the two major volcanoes which constitute the island (Stearns and Vaksvik, 1935). Its lavas, known as the Waianae volcanic series, are divided into the lower, middle, and upper members (Macdonald, 1940a). The lower member consists of the innumerable thin lava flows constituting the broad shield volcano which makes up the major part of the mountain. Its lavas are all basaltic. Olivine basalt is by far the most common type, although olivine-poor basalts, olivine-free basalts, and picrite-basalts of the oceanite type also are present. A few flows contain hypersthene. Phenocrysts of plagioclase and pyroxene are rare.

The middle member of the Waianae volcanic series comprises the lavas erupted into the caldera of the Waianae Volcano. For the most part its lavas closely resemble those of the lower member, but plagioclase and pyroxene phenocrysts are more abundant, and olivine-poor basalts are more numerous. Some of the latter

are transitional toward andesites. A few basaltic andesites are present in the upper part of the middle member. Mauna Kuwale, a small peak between Waianae and Lualualei valleys, consists largely of trachyte, underlain by basalts and overlain by basaltic andesite. It was previously suggested (Macdonald, 1940a) that the trachyte belongs with the upper member of the Waianae volcanic series, and is situated topographically below the base of the upper member because it accumulated in a pit crater. Subsequent careful search for any evidence of faulting, or other indication of the hypothetical pit crater, has been entirely fruitless. It appears better to regard the trachyte as erupted in the caldera of Waianae Volcano toward the end of the caldera-filling period. It thus corresponds approximately in stratigraphic position to the trachyte of Puu Waawaa on Hawaii, and to at least some of the trachytes of Tutuila, American Samoa (Macdonald, 1944e).

The upper member of the Waianae volcanic series consists predominantly of thick massive flows of basaltic andesite, although both olivine basalts and olivine-free basalts are intercalated in its lower part. A few inclusions of dunite and gabbro are found. Late cinder cones of the upper member have been stated to consist of nepheline basalt and basanite (Cross, 1915, p. 23; Powers, 1920, p. 273), but none of them is devoid of plagioclase, and staining has revealed no nepheline. Moreover, a chemical analysis of one of them shows the presence of only 1.4 per cent of normative nepheline. It is therefore believed that nepheline basalt and basanite are absent on Waianae Volcano.

After a long period of erosion during which the Waianae Volcano was largely if not entirely quiescent, a single small flow of olivine basalt (Stearns, 1940, p. 47) was erupted.

Koolau Volcano.—Koolau Volcano forms the eastern part of the island of Oahu. It is a much-dissected elongate shield volcano, composed very largely of olivine basalt, with less abundant flows of olivine-poor basalt and picrite-basalt of the oceanite type. Many of the rocks contain small phenocrysts of hypersthene (Wentworth and Winchell, 1947). In petrographic character, the lavas of the Koolau Volcano are much like those of Mauna Loa on the island of Hawaii.

On the eastern slope of the Koolau Range there is an area of thick massive lava flows (the Kailua volcanic series) which are regarded as the caldera-filling lavas of the Koolau Caldera (Stearns, 1940, p. 48–50). The lavas have been extensively affected by rising volatiles, with the formation of veinlets and amygdules of quartz and chalcedony. The ferromagnesian minerals and the glassy portions of the groundmass have been partly altered to chlorite. Similar alteration is encountered in the lavas occupying the caldera of East Molokai Volcano.

Honolulu volcanic series.—A long period of erosion followed extinction of the Koolau Volcano, and great valleys were carved into the mountain, some of them more than 1000 feet deep. After this long interlude of quiet, volcanism was renewed with the eruption of a series of lava flows and pyroclastic cones which have been named the Honolulu volcanic series (Stearns and Vaksvik, 1935). The lavas of the Honolulu volcanic series have been studied in detail by Horace Winchell (1947). They include nepheline basalt, melilite-nepheline basalt, nepheline basanite, and "linosaite". Winchell believes that they can be divided into several groups, each representing the injection of magma into a rift zone, and within each of which the succession of extrusion was first very slightly feldspathoidal olivine basalt ("linosaite") or basanite, and later nepheline basalt. Winchell has identified at least small amounts of nepheline in all the flows, with the possible exception of the Kaupo flow (which may be the latest flow of the series). It therefore appears that the Kaupo, Black Point, Kaohikaipu, Koko, Mauumae, and Kaimuki volcanics, reported by Stearns and Vaksvik (1935, p. 196) as olivine basalt, are weakly feldspathoidal and are to be grouped with the so-called "linosaites."

Molokai

West Molokai Volcano.—West Molokai consists of a little-dissected shield volcano, the eastern side of which has been partly downfaulted (Fig. 5). There is no evidence of the existence of a caldera. The rocks are predominantly olivine basalts, with a smaller proportion of olivine-poor and olivine-free basalts, and picrite-basalts of the oceanite type (Macdonald, 1947a). In degree of variation, the lavas resemble those of Kilauea, Mauna Loa, the

FIGURE 5.—GEOLOGIC MAP OF THE ISLAND OF MOLOKAI
Simplified after Stearns and Macdonald (1947, Pl. 1).

Koolau Volcano on Oahu, and the older parts of the Waianae, Mauna Kea, Kohala, and Haleakala volcanoes. Many of the lavas are nonporphyritic and exceptionally thin-bedded, probably indicating high magmatic temperature during eruption. In the zone of faulting on the eastern side of the shield are innumerable nodules of chalcedony, quartz, and calcite, probably deposited by rising deep-seated volatiles. Most of the latest lavas of West Molokai are olivine basalts, but two flows are basalt poor in olivine. These late lavas were extruded after a period of weathering long enough to develop 3 or 4 feet of residual soil on the earlier lavas.

East Molokai Volcano.—The eastern part of the island of Molokai is essentially a broad basaltic shield volcano, cut on the north by a high cliff which may be an eroded fault scarp, and dissected by huge erosional valleys. The lavas are predominantly olivine basalt, with less abundant olivine-poor and olivine-free basalt and picrite-basalt. In the lower part of the section the picrite-basalts are of the oceanite type, but in the upper part picrite-basalts of the ankaramite type are common. Within the area formerly occupied by the caldera of the volcano, the rocks are much altered by rising volatiles. Ferromagnesian minerals and glassy groundmass material are partly or largely changed to chlorite; and quartz, chalcedony, and zeolites have been deposited in vesicles and other open spaces. Outside the caldera areas of the Hawaiian volcanoes amygdules are rare, but within both the East Molokai and Koolau (Oahu) calderas amygdaloidal rocks are very common. Small stocks of gabbro, closely resembling those of Kauai and West Maui, intrude the basalts of East Molokai.

A relatively thin but extensive cover of less mafic rocks overlies the basaltic lavas of East Molokai. Most of the flows are oligoclase andesite, but a few are andesine andesite. Trachyte has been found in place only at the dome of Puu Kaeo, on the western edge of Waikolu Valley, and the small cinder cone of Puu Anoano on the northwestern slope of the mountain; however, pebbles of trachyte reported by Powers (1920, p. 271) in Wailau Stream prove the occurrence of trachyte somewhere in the upper drainage of that stream.

Following a long period of erosion lavas were erupted at the foot of the great northward-facing cliff, building a small shield volcano which projects a short distance above sea level as the Kalaupapa Peninsula. The lavas are mafic olivine basalt. At about the same time similar magma rising beneath the ocean a short distance east of Molokai exploded on contact with sea water and built the cone of palagonitized tuff which now forms Mokuhooniki (Hooniki Island).

Lanai

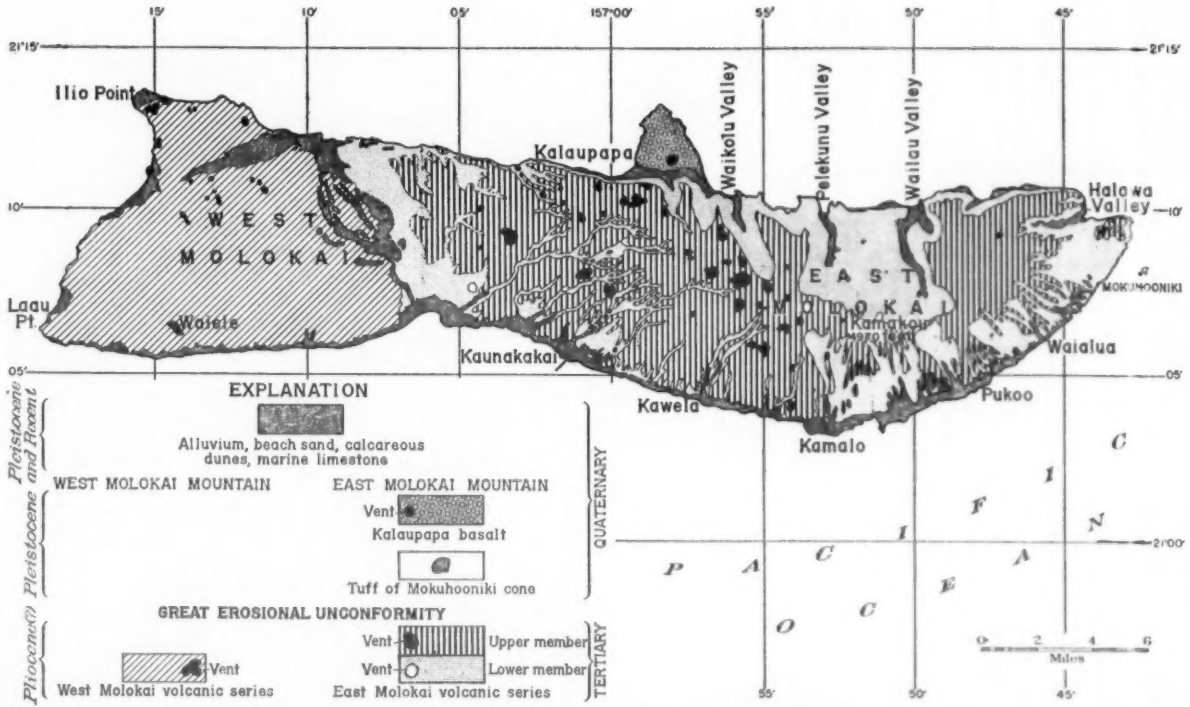
The island of Lanai is a basaltic shield volcano. A large caldera, at or near the summit of the shield, has been largely filled with later lavas. The rocks appear to be all basaltic. All of the specimens examined are olivine basalt or basalt, although a few show a tendency toward the basaltic andesites (Macdonald, 1940b). Small phenocrysts of hypersthene occur in one specimen. A few flows of picrite-basalt of the oceanite type are present.

Kahoolawe

Kahoolawe Island is a basaltic shield volcano in which a caldera developed and then was filled and overflowed by later lavas. The pre-caldera lavas are principally thin flows of olivine basalt. Less abundant are olivine-poor and olivine-free basalts. A few flows contain small phenocrysts of hypersthene (Macdonald, 1940c). Other rock types have not been found in the pre-caldera lavas, which are exposed only in part of the sea cliffs along the eastern and southern shores, but further search probably would reveal picrite-basalts of the oceanite type.

Most of the earlier caldera-filling lavas resemble the pre-caldera flows. Interbedded with these, however, and becoming more abundant in the upper part of the section, are basaltic andesites. Eventually, the caldera was filled, and there was built over it, and along the rift zone to the west, broad gently sloping cones of lavas ranging from olivine basalt to andesite. Basaltic andesite is the most abundant type among the post-caldera lavas.

Following the cessation of post-caldera volcanism, marine erosion developed a high sea cliff along the eastern and southern shores of Kahoolawe. Finally, after a long period of quiescence, at five points on the sea cliff on the



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eastern side of the island, there were small eruptions of olivine basalt.

Maui

West Maui Volcano.—The lavas of West Maui Volcano (Fig. 6) are divided into the Wailuku, Honolua, and Lahaina volcanic series (Stearns and Macdonald, 1942). The Wailuku volcanic series comprises the lavas which built the broad shield volcano which constitutes most of the bulk of West Maui. The lavas are preponderantly olivine basalt, grading on the one hand into olivine-poor basalt and on the other into picrite-basalt of the oceanite type. Small phenocrysts of hypersthene have been found in some of the olivine basalts and in one picrite-basalt (Macdonald, 1942b, p. 312–334). Dunite inclusions occur in a few of the flows. Several small stocks of gabbro intrude the Wailuku lavas.

The Honolua volcanic series consists largely of oligoclase andesite, with some trachyte. The oligoclase andesites form a thin veneer over much of the surface of the original basaltic shield. The trachytes form thicker, more localized flows, and steep-sided domes (tholoids). There is no gradation from the basalts of the Wailuku volcanic series to the oligoclase andesites of the Honolua volcanic series.

Cessation of volcanism and extensive erosion of the West Maui Volcano was followed by a few small eruptions, building the cinder cones and lava flows of the Lahaina volcanic series. Most of the Lahaina lavas are picrite-basalts of the ankaramite type, but one is a nepheline basanite.

Haleakala Volcano. — Haleakala Volcano forms the eastern part of the island of Maui. The oldest rocks comprise the Honomanu volcanic series, which is exposed only along a short section of the sea cliff and in some of the deeper valleys. All of the Honomanu lavas studied are olivine basalts, although some in the uppermost part of the section contain augite phenocrysts and are transitional toward the augite-rich (ankaramite) type of picrite-basalt. Others, also in the uppermost part, are transitional toward the andesites in feldspar composition. Many olivine basalts of the Honomanu volcanic series contain numerous large phenocrysts of feldspar. Although neither type has been found, the Honomanu lavas probably include at least a few flows of olivine-poor basalt and picrite-basalt of the oceanite type.

The Honomanu lavas grade into the overlying Kula volcanic series. The most abundant lavas of the Kula volcanic series are basaltic andesites, but many flows of olivine basalt and augite-rich picrite-basalt, and some flows of andesite, also are present. There is a complete gradation from basaltic andesite through typical andesine andesite to a few examples of oligoclase andesite.

The latest, or Hana volcanic series in many places is separated from the Kula volcanic series by a profound erosional unconformity. Olivine basalts are more numerous among the Hana lavas than among the Kula lavas. Basaltic andesites and picrite-basalts of the ankaramite type are abundant, but typical andesites have not been found. The latest lava flow from Haleakala Volcano, which occurred about 1750, is augite-rich picrite-basalt of the ankaramite type.

The islet of Molokini, between Maui and Kahoolawe islands, is a tuff cone probably formed by a submarine eruption on the southwest rift zone of Haleakala Volcano. The fresh tuff consists of black glassy fragments of olivine basalt, but much of it is altered to yellowish-brown palagonite (Palmer, 1930; Macdonald, 1942b, p. 305).

Hawaii

Kohala Volcano.—Kohala, the northernmost and oldest of the five major volcanoes which have built the island of Hawaii (Fig. 7), is an elongate basaltic shield volcano, partly veneered by flows of andesite and trachyte. The older lavas comprise the Pololu volcanic series (Stearns and Macdonald, 1946), and are nearly all olivine basalts. Only a very small proportion are olivine-poor basalts, and no olivine-free basalts or picrite-basalts have been found. Some of the olivine basalts contain very abundant large feldspar phenocrysts. Small stocks of olivine gabbro intrude the Pololu lavas.

The later lavas, comprising the Hawi volcanic series, veneer much of the western slope of the mountain, but fault scarps parallel to the rift zone permitted the escape eastward of only a few of the latest flows. The Hawi lavas are nearly all oligoclase andesites, but a few examples of andesine andesite are known (Mac-

FIGURE 6.—GEOLOGIC MAP OF THE ISLAND OF MAUI
Simplified after Stearns and Macdonald (1942, Pl. 1).

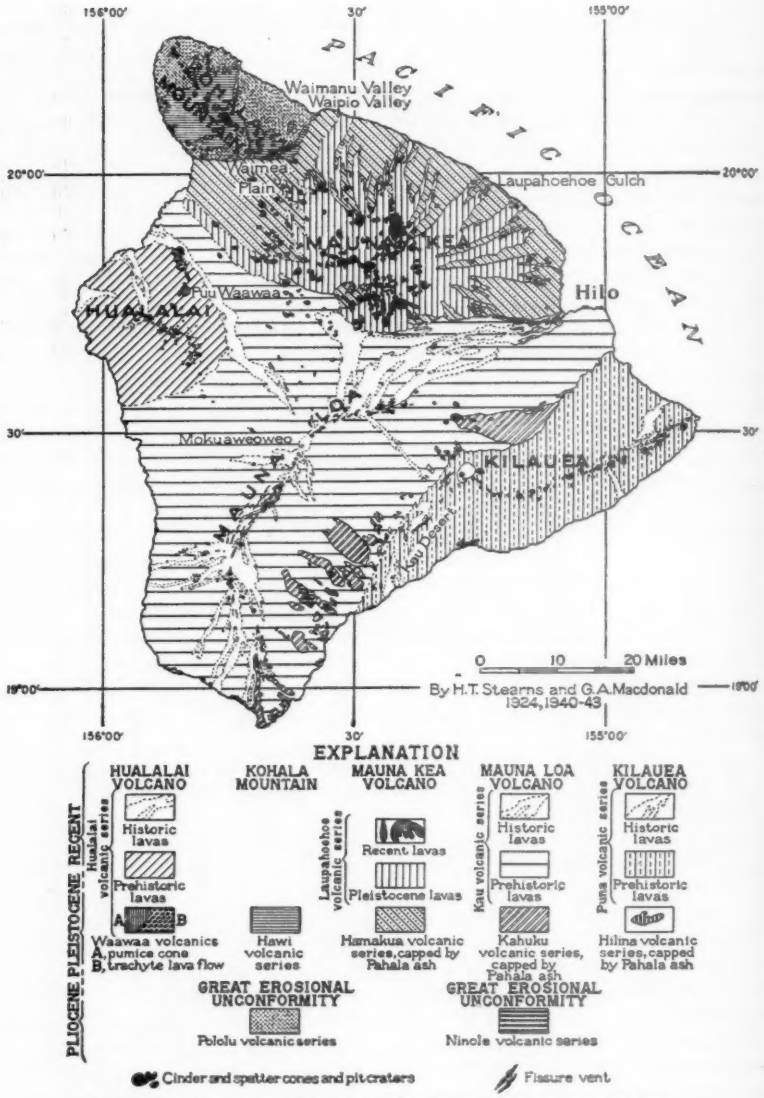

FIGURE 7.—SIMPLIFIED GEOLOGIC MAP OF THE ISLAND OF HAWAII
After Stearns and Macdonald (1946, Fig. 17).

donald, 1946). Trachyte has been found in one dike and in several small viscous domes with associated short thick lava flows. Xenoliths in the lavas include dunite, olivine gabbro, and augite-hypersthene peridotite.

Mauna Kea.—Mauna Kea is the second oldest volcanic mountain of the island of Hawaii in time of extinction. Its lower slopes exhibit the broad, nearly flat profile typical of basaltic shield volcanoes. The upper slopes are much steeper, however, and approach the profile typical of andesitic volcanoes of continental

areas. The greater steepness of the upper slopes results partly from a greater abundance of pyroclastic material and partly from the presence of andesite flows which were erupted in a more viscous condition than the basalts and tended to accumulate closer to the vents. The rocks of Mauna Kea are divided into the Hamakua and Laupahoehoe volcanic series. The Hamakua volcanic series, which is the older of the two, may be subdivided into a lower and an upper member. The lower member consists largely of olivine basalt, with which are associated lesser amounts of olivine-poor basalt and picrite-basalt of the oceanite type. The lower member grades into the upper member, in which olivine basalt and olivine-poor basalt are likewise present, but are interbedded with picrite-basalt of the ankaramite type and andesine andesite. Transitions from olivine basalt to the ankaramite type of picrite-basalt and to basaltic andesite are present, just as they are on Haleakala Volcano.

The rocks of the Laupahoehoe volcanic series are preponderantly andesine andesite, but basaltic andesite and olivine basalt also are present. No picrite-basalt has been found. The transition from the Hamakua to the Laupahoehoe volcanic series is gradual. The lavas and pyroclastic rocks of the Laupahoehoe volcanic series contain many fragments of gabbro, dunite, and augite periodotite. Six very recent lava flows near the summit of Mauna Kea, probably erupted since the Pleistocene glaciation of the mountain, are all andesine andesite.

Hualalai Volcano.—The lavas of Hualalai are very largely olivine basalts, but there also occur lavas transitional to the picrite-basalts of ankaramite type, olivine-poor basalts, a few basaltic andesites, and at least one flow of andesine andesite. No true picrite-basalts have been found. Puu Waawaa, on the northwestern flank of the mountain, is a large trachyte cinder cone northwestward from which extends a thick flow of trachyte (Cross, 1904). Puu Anahulu is a small hill on the surface of the trachyte flow. The trachyte was both preceded and followed by flows of olivine basalt from Hualalai and is partly buried by flows from Mauna Loa. Xenoliths in the lavas of Hualalai include olivine gabbro, dunite, augite periodotite, and intergradational types. The latest lava, erupted in 1801, is an olivine basalt.

Mauna Loa.—The lavas of Mauna Loa have been divided into three groups, of different age (Stearns and Clark, 1930, p. 61; Stearns and Macdonald, 1946). The oldest is the Ninole volcanic series, exposed in a small area on the southeastern side of the ridge which marks the southwest rift zone of the volcano. The lavas dip not from the summit area of Mauna Loa but from an area roughly 18 miles to the southwest, in the vicinity of Puu o Keokeo, and it was suggested by Stearns and Clark that they belong to a volcanic center older than Mauna Loa, and situated farther southwest. Bishop (1907) had previously suggested the existence of such an independent volcano in the neighborhood of Puu o Keokeo. However, later lavas definitely belonging to Mauna Loa exhibit similar dips, projecting to points along the rift zone. Although the possibility of an older volcano southwest of the present Mauna Loa, and largely buried by it, cannot be disproved, there appears to be little or no evidence in its favor.

The Ninole lavas were deeply eroded, with the formation of large stream-cut valleys, and the next succeeding lavas overlie them with profound unconformity. The second group of lavas are included in the Kahuku volcanic series. At its top, a thick bed of ash (Pahala ash) separates its lavas from those of the succeeding Kau volcanic series. The eruption of the Kau lavas has continued uninterrupted to the present time.

The lavas of all three groups are essentially similar. All consist principally of olivine basalt, with less abundant basalt and picrite-basalt of the oceanite type. Many lavas of all three groups contain small phenocrysts of hypersthene, and among the historic lavas the proportion of flows containing hypersthene rises to about 47 per cent. The proportion of olivine basalt is greater in the Ninole and Kahuku rocks than in those of the Kau volcanic series. Basalts free of olivine occur in the Kau volcanic series but have not been found among the earlier lavas. Some of the basalts of the Kau volcanic series, both with and without olivine, approach the andesites in the average composition of their feldspar, but no rocks which can be classed as andesites, even basaltic andesites, have been found. Table 2 summarizes the occurrence of the various types among the lavas of Mauna Loa and Kilauea.

Kilauea Volcano.—The lavas of Kilauea are largely comprised in the Puna volcanic series. Flows exposed beneath the Puna lavas in small windows along fault scarps south of Kilauea

TABLE 2.—PROPORTIONS OF THE VARIOUS ROCK TYPES AMONG LAVAS OF KILAUEA AND MAUNA LOA VOLCANOES

Volcano	Stratigraphic unit		Olivine basalt	Basalt	Hypersthene-bearing basalt	Picrite-basalt (oceanite type)
Kilauea	Puna volcanic series	Historic	78	13	2	7
		Prehistoric	71	22	2	5
	Hilina volcanic series		67	30	0	3
Mauna Loa	Kau volcanic series	Historic	16	26	47	11
		Prehistoric	43	26	10	21
	Kahuku volcanic series		60	22	8	10
	Ninole volcanic series		62	15	8	15

Caldera have been named the Hilina volcanic series (Stearns and Macdonald, 1946). They formerly were mapped as part of the "Pahala basalt" (Stearns and Clark, 1930) and are probably essentially contemporaneous with the Kahuku lavas on Mauna Loa. It is uncertain whether they originated from Kilauea or from Mauna Loa. The lavas of both the Puna and the Hilina volcanic series are preponderantly olivine basalts, although basalts and picrite-basalts of the oceanite type also are present. A few of the lavas of the Puna volcanic series contain hypersthene, but that mineral is much less common than in the lavas of Mauna Loa. Kilauean lavas show less variation than those of Mauna Loa.

Summary of Stratigraphic Relations

The great bulk of each of the major Hawaiian volcanoes consists of olivine basalt. This rock type probably constitutes 65 per cent or more of the visible islands and an even larger proportion of the immensely greater bulk beneath sea level. In the primitive shield volcano which forms the major part of each mountain, there is associated with the olivine basalt a small proportion of picrite-basalt of the oceanite type, and of basalt nearly or entirely free of olivine. In the uppermost part of the primitive shield, picrite-basalts of the ankaramite type make their appearance.

Some of the major Hawaiian volcanoes have not passed beyond the stage of the primitive shield volcano. Two of these, Mauna Loa and Kilauea, are still active. Two others, Lanai and West Molokai, appear to be extinct. The Koolau Volcano on Oahu did not pass beyond the primitive stage during the principal period of volcanism, although it produced highly differentiated nepheline basalts at a much later date. Niihau also did not pass beyond the primitive stage during the principal eruptive period, and even the post-erosional lavas are olivine basalt. The other Hawaiian volcanoes advanced beyond the stage of the primitive basaltic shield and during the latter part of the principal period of volcanism erupted andesites or even trachytes.

The volcanoes which erupted andesites may be divided into two general types. In one, the passage upward from olivine basalts to andesites is gradual. Andesites and basaltic rocks are interbedded, and rocks intermediate in composition between basalts and andesites occur. Intercalated with them are picrite-basalts of the ankaramite type. The andesites themselves are very largely andesine andesites, although minor amounts of oligoclase andesite are found. Olivine basalt recurs throughout the section. This type is well exemplified by Haleakala and Mauna Kea volcanoes. Trachyte may occur as small bodies intercalated with the basaltic and andesitic flows, as at Hualalai and Waianae volcanoes.

In contrast, in the other type there is no gradation from the primitive shield to the andesite capping. The andesites are oligoclase andesites, associated with and apparently grading into trachytes. The trachytes appear in general to be later than most of the andesites. There is no recurrence of basaltic rocks in the andesitic part of the section, and the andesites often are separated from the basalts by a period of deep weathering and erosion. This type is exemplified by West Maui and Kohala volcanoes.

Following a long period of volcanic quiescence and deep erosion, several of the Hawaiian volcanoes have renewed their activity. At Haleakala Volcano the new lavas were much like those extruded just previous to the quiet period, but with a greater proportion of basaltic rocks. On the southern slope of Mauna Loa a similar break appears to exist between the Ninole volcanic series and the later rocks. At the East Molokai and Niihau volcanoes the post-erosional lavas are ordinary olivine basalt, or olivine basalt transitional to picrite-basalt. At West Maui Volcano, the Koolau Volcano on Oahu, and Kauai Volcano, the long erosional period was followed by eruption of basic and ultrabasic alkaline rocks (basanites, nepheline basalts, and melilite-nepheline basalts), together with olivine basalts similar to those of the primitive shield volcano, and on Kauai the mimosite type of picrite-basalt.

Table 3 summarizes the distribution of the principal rock types in the Hawaiian Islands.

Petrogenesis

General Statement

Much is now known of the composition and stratigraphic relationships of the types of rock composing the Hawaiian Islands. Indeed, few volcanic districts are known more thoroughly. In view of that, it appears both justifiable and desirable to apply to the province certain of the more widely held theories of petrogenesis; and by testing them against the known facts to either accept them, reject them, or so modify them that they furnish as complete and reasonable a picture as can now be obtained of the origin of the igneous rocks of the Hawaiian province. Such an analysis is here attempted. The resulting picture is tentative and incomplete, and there is no intention to extrapolate it beyond the Hawaiian province and similar mid-oceanic provinces. Further research in this and other provinces certainly will bring forth new theories and better understanding of the petrogenic processes involved. It is believed, however, that the picture is the most nearly complete and most accurate of any thus far advanced for the Hawaiian province.

Parent Magma

The composition of the undifferentiated magma of the Hawaiian province cannot be stated precisely, but there appears to be little question that it approximates olivine basalt. Olivine basalt magma is not, of course, the same in composition as the solidified olivine basalt, as the process of solidification involves the loss of large amounts of volatile material. The gases liberated at Kilauea have been collected and analyzed (Jaggar, 1940), and the general nature of the gases and their approximate relative proportions are known. However, one can formulate only a general estimate as to the proportion of these gases dissolved in the magma beneath the surface of the earth, or even the proportion carried by a lava flow erupted onto the surface. It is necessary largely to ignore the volatiles in considering the composition of the parent magma and the steps by which the other rock types were derived from it.

The very great predominance of olivine basalt among the early visible lavas of the Hawaiian volcanoes constitutes strong evidence that olivine basalt represents the nonvolatile fraction of the primary Hawaiian magma. This conclusion is further strengthened by the fact that all the other rock types can be derived from such a magma by relatively simple and logical processes. The parent magma of the Hawaiian province may safely be assumed to be olivine basalt. It is necessary to go further, however, and attempt to arrive at the approximate chemical composition of the nonvolatile phase of this magma.

Cross (1915, p. 87) used as this composition an average (Table 4, column 6) of all existing analyses of Hawaiian rocks, and Cross' average has recently been used by Winchell (1947). A similar average based on the much more numerous analyses now available is shown in column 7 of Table 4. The use of such an average is open to serious criticism because: (1) Although the magma may differentiate into complementary fractions, it does not follow that these fractions will be represented in their true relative abundance by surface outflows. Indeed, it appears almost a foregone conclusion that the felsic fraction, which rises to a high level in the magma reservoir, will be erupted more abundantly than the mafic fraction, which sinks in the reservoir. (2) The chemical analyses of the various types of rock are in far different abundance than the actual types in nature. The basaltic rocks, particularly the olivine basalts,

TABLE 3.—DISTRIBUTION OF PRINCIPAL TYPES OF LAVAS IN THE HAWAIIAN ISLANDS

Types of lavas	Island and volcano																													
	Laysan	Gardner	French Frigate	Necker	Nihoa	Kaula	Lehua	Niihau	Kauai		Oahu		Molokai		Lanai	Maui		Hawaii											
									Kauai Volcano	Haupu Volcano	Waianae Volcano	Koolau Volcano	West Molokai Volcano	East Molokai Volcano		West Maui Volcano	Haleakala Volcano	Kohala Volcano	Mauna Kea Volcano	Hualalai Volcano	Mauna Loa Volcano	Kilauea Volcano	Molokini	Kahoolawe
Olivine basalt	X	X	X	X	X	*	*	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	*	X
Basalt		X			X	*		X	X	X	X	X	X	X	X	X	X	X	X	X	X	X		X
Hypersthene-bearing basalt									X		X	X	X		X	X					X	X		X
Picrite-basalt, oceanite type					X			X	X	X	X	X	X	X	X	X	X		X		X	X		
Picrite-basalt, ankaramite type				X										X		X								
Picrite-basalt, mimosite type									X															
Andesine andesite				X	X			X	X		X			X		X	X	X	X	X	X	X		X
Oligoclase andesite													X	X		X	X							
Trachyte											X		X	X		X	X			X				
Nepheline basanite				X					X			X				X								
Nepheline basalt									X			X												
Melilite-nepheline basalt									X			X												
Palagonite tuff						X	X					X		X				X				X	X	
Gabbro { intrusives									X		X	X		X		X		X	X	X	X			
Gabbro { inclusions		X							X								X	X	X	X	X			
Dunite inclusions					X				X		X	X		X		X	X	X	X	X	X			
Peridotite inclusions†												X		X		X	X	X	X	X	X			

* Present as ejected blocks only.

† Other than dunite.

are greatly underemphasized in the collections, and the less abundant andesites, trachytes, and nepheline basalts are greatly overemphasized. To some extent the analyses of the salic andesites and trachytes and the ultramafic nepheline basalts tend to compensate each other, so that in some respects the average approaches closely the composition of olivine basalt. But both the trachytes and the nepheline basalts are richer in alkalies than is olivine basalt, and the average of all available chemical analyses is decidedly higher in alkalies than olivine basalt. It is much simpler to derive the nepheline basalts from such a high-alkali magma than from olivine basalt magma, but the assumption of such a magma as the parent in the Hawaiian province appears to the writer wholly unjustified.

In an earlier paper dealing with the island of Hawaii (Macdonald, 1949), the writer assumed as the parent magma the average of all existing first-class analyses of rocks of Kilauea Volcano, on the grounds that Kilauean lavas show relatively little variation in composition and are represented by more analyses than the lavas of any other Hawaiian volcano. The average is quite similar to that for the lavas of Mauna Loa, which also show relatively little variation, and to the average of Hawaiian olivine basalts. Its use is subject to the same criticisms as is the use of the general average of all Hawaiian rocks, though to a much lesser degree. The variations in Kilauean lavas are small and consist largely in variations in olivine content. The analyses of olivine-poor and olivine-free basalt are to some extent counteracted by less numerous analyses of olivine-rich picrite-basalt.

Daly (1944) has assumed olivine basalt to be the parent magma of the Hawaiian province. His average analysis (Table 4, column 4) contains, however, besides olivine basalt, several analyses of olivine-poor or olivine-free basalt.

In the present paper the nonvolatile fraction of the parent magma has been assumed to have approximately the composition shown in column

TABLE 4.—AVERAGE ANALYSES OF GROUPS OF HAWAIIAN LAVAS

	1	2	3	4	5	6	7	8	9	10	11
SiO_2	48.35	49.80	50.42	50.04	50.18	48.69	48.03	49.06	48.46	45.	50.
Al_2O_3	13.18	12.42	11.62	13.47	14.12	14.00	13.48	15.70	13.80	15.	13.
Fe_2O_3	2.35	1.53	2.71	2.65	3.04	5.03	3.48	5.38	3.35	13.	13.
FeO	9.08	9.91	9.07	9.05	10.45	8.01	8.15	6.37	9.96		
MgO	9.72	10.31	10.11	7.79	6.42	7.12	8.84	6.17	6.25	8.	5.
CaO	10.34	10.32	9.74	10.49	10.67	9.00	9.53	8.85	9.55	9.	10.
Na_2O	2.42	1.96	2.09	2.43	2.69	3.55	3.19	3.11	2.84	2.5	2.8
K_2O	0.58	0.45	0.39	0.55	0.75	1.24	0.96	1.52	0.98	0.5	1.2
TiO_2	2.77	2.68	2.97	3.12	1.75	2.29	2.87	1.36	2.55	—	—
P_2O_5	0.34	0.29	0.27	0.29	0.21	0.49	0.47	0.45	0.46	—	—
MnO	0.14	0.13	0.10	0.12	0.19	0.41	0.15	0.31	0.21	—	—

1. Olivine basalt of the Hawaiian Archipelago; average of 53 analyses, including one of olivine gabbro.
2. Lava of Kilauea; average of 24 analyses.
3. Lava of Mauna Loa; average of 14 analyses.
4. Average of 23 analyses of Hawaiian olivine basalts and basalts (Daly, 1944, p. 1370).
5. Average of 9 historic lavas of Kilauea, belonging to the camptonose-ornose-auvergnose group, which Cross found to be the most abundant group, on the basis of the C. I. P. W. classification, in the Hawaiian Islands (Cross, 1915, p. 87).
6. Hawaiian lava; average of 43 analyses of all types of lava from the whole Hawaiian Archipelago (Cross, 1915, p. 87).
7. Hawaiian lava; average of 151 analyses of all types of lava from the whole Hawaiian Archipelago.
8. Average composition of world basalts (Daly, 1933, p. 17).
9. Plateau basalt; average of 50 analyses from the Deccan, Columbia River, and Thulean (North Atlantic) regions. Data from Washington (1922).
10. Olivine basalt magma type of Kennedy and Anderson (1938, p. 36).
11. Tholeiitic magma type of Kennedy and Anderson (1938, p. 36).

1 of Table 4, which is the average of 53 analyses olivine basalt from many parts of the Hawaiian Archipelago.

The Hawaiian averages are compared in Table 4 with the average composition of world basalts (column 8) and the average plateau basalt (column 9). The average Hawaiian olivine basalt is somewhat more femic than the average world basalt. It is surprisingly close to the average plateau basalt, although it contains a little more magnesia and lime, and a little less alumina, iron, and alkalies. Columns 10 and 11 show respectively the compositions of Kennedy's hypothetical olivine basalt and tholeiitic magma types (Kennedy and Anderson, 1938, p. 36). The average Hawaiian olivine basalt resembles Kennedy's olivine basalt magma type in the proportions of the alkalies, and is even richer than it in magnesia; but more closely resembles his tholeiitic magma type in the percentages of silica, alumina, and lime.

Chemical Composition of Major Rock Types

Table 5 lists the average chemical composition and norm of each principal rock type of the Hawaiian Islands. Nearly all the modern chemical analyses of Hawaiian lavas used in the present work, have been published previously, some of them several times, in readily available sources (Daly, 1911; Cross, 1915; Washington, 1923a; 1923b; 1923c; Washington and Keyes, 1926; 1928; Piggot, 1931; Powers, 1931; Shepherd, 1938; Macdonald, 1940a; 1942b; 1946; 1947a; 1949; Macdonald and Powers, 1946; Wentworth and Winchell, 1947; Winchell, 1947). There appears to be no need to repeat them here, except as they form the basis of special discussions. Three hitherto unpublished analyses are included in columns 2, 4, and 5 of Table 6. One analysis of a rock from Necker Island (Washington and Keyes, 1926, p. 347, no. 3) has been omitted from the average of picrite-basalts of ankaramite type, because it is markedly richer in potassia and lower in alumina than the other analyses of rocks of that type and may not actually belong with them.

The variation diagram, Figure 8, has been prepared from the 151 available modern analyses of Hawaiian rocks. Several older analyses have been omitted because they are not directly comparable with the more modern analyses. The diagram is constructed by the method devised by Larsen (1938), in which the percentage of the various oxides are plotted as the ordinate against the value ($\frac{1}{3}$ SiO_2 + K_2O) minus (FeO + MgO + CaO) on the abscissa. This method gave a much more nearly linear distribution of points than the more common method of plotting the percentages of the other oxides against that of silica.

Because of the wide scattering of values for the oxides involved, particularly soda, the alkali-lime index cannot be stated with accuracy. However, it is in the vicinity of 54, placing the Hawaiian province as a whole in the alkali-calcic series (Peacock, 1931), although certain individual Hawaiian volcanoes such as Haleakala appear to belong in the alkaline series (Macdonald and Powers, 1946). It should be noted that the alkali-lime index is determined by the rocks more silicic than olivine basalt. Although soda is abundant in some of the rocks less silicic than olivine basalt, in none of them does the sum of the alkalies equal the proportion of lime. In these mafic rocks the curve for soda (Fig. 8) appears to split into two moderately distinct limbs, the upper of which represents the nepheline-bearing lavas. This part of the series is distinctly alkalic. A similar splitting into two more or less distinct limbs is evident near the left, or mafic, end of the curves for most of the other oxides, the nepheline-bearing rocks being predominant in one limb and absent in the other.

Order of Crystallization in Olivine Basalt

The observed order of crystallization in Hawaiian olivine basalts is approximately as shown in Figure 9. Essentially, the phenocrysts represent intratelluric crystallization, and the groundmass represents crystallization after extrusion. The earliest phenocrysts to form are olivine, which in many rocks at first separates in excess of its stoichiometric proportion and later is partly resorbed. Rarely, zoning is detectable in the olivine, the outer layers being richer in iron than the core. At and just preceding the time of eruption, concentration of gas in the upper part of the magma column may cause partial alteration of the olivine to iddingsite, followed by renewed precipitation of olivine during the period of crystal-

TABLE 5.—AVERAGE COMPOSITION OF PRINCIPAL ROCK TYPES

ANALYSES

	1	2	3	4	5	6	7	8	9	10	11	12	13
SiO_2	36.73	38.66	39.44	41.26	43.36	45.00	45.30	46.74	48.35	48.76	50.44	51.35	62.77
Al_2O_3	10.78	10.85	10.24	10.31	11.41	13.00	11.76	8.19	13.18	15.82	14.00	16.34	17.20
Fe_2O_3	5.57	5.82	6.52	5.58	2.82	4.04	3.98	2.77	2.35	4.10	3.15	4.64	3.14
FeO	8.76	7.75	7.02	8.62	10.17	8.32	8.80	9.60	9.08	7.53	8.27	6.19	1.30
MgO	12.74	13.56	14.14	14.00	14.61	9.03	12.88	20.47	9.72	4.74	6.73	3.73	0.69
CaO	13.70	13.04	12.32	11.65	11.05	10.16	10.72	7.47	10.34	7.99	10.19	6.61	1.43
Na_2O	3.88	4.00	2.67	3.19	1.75	4.75	2.36	1.74	2.42	4.50	2.58	5.01	6.74
K_2O	0.91	1.12	1.21	0.88	0.59	1.18	0.72	0.35	0.58	1.58	0.48	1.94	4.11
TiO_2	2.84	2.71	3.30	2.66	3.04	3.16	2.30	2.12	2.77	3.29	3.06	2.74	0.53
P_2O_5	1.10	1.03	0.78	0.64	0.34	0.48	0.31	0.22	0.34	0.72	0.29	1.00	0.21
MnO	0.12	0.10	0.09	0.14	0.16	0.11	0.11	0.14	0.14	0.17	0.12	0.20	0.19

NORMS

Q	—	—	—	—	—	—	—	—	—	—	4.14	—	4.32
or	—	—	—	—	3.34	7.27	3.89	2.22	3.34	9.45	2.78	11.12	24.46
ab	—	—	—	—	11.53	4.72	16.77	14.67	20.44	30.39	22.01	42.44	57.11
an	9.17	8.06	12.23	10.84	21.68	10.29	19.74	13.34	23.35	18.07	25.02	16.40	4.45
ne	17.89	18.46	12.21	14.77	1.70	19.31	1.70	—	—	4.26	—	—	—
lc	4.36	5.23	5.67	4.36	—	—	—	—	—	—	—	—	—
cs	7.22	4.47	—	—	—	—	—	—	—	—	—	—	—
di wo	11.83	14.96	18.10	18.10	12.64	15.20	13.22	9.51	10.90	9.51	9.86	4.18	0.81
di en	8.70	11.60	15.00	13.60	8.80	10.70	9.40	7.00	7.10	6.20	6.30	2.80	0.70
di fs	1.98	1.72	0.79	2.64	2.77	3.17	2.64	1.58	3.04	2.64	2.90	1.06	—
hy en	—	—	—	5.40	—	—	—	11.30	10.00	—	10.30	2.90	1.00
hy fs	—	—	—	1.06	—	—	—	2.64	4.22	—	4.62	1.06	—
ol fo	16.10	15.61	14.28	11.20	19.32	8.33	15.96	23.03	5.04	3.92	—	2.52	—
ol fa	3.88	2.55	1.02	2.45	6.73	2.75	5.00	6.02	2.45	2.04	—	1.02	—
mt	8.12	8.35	9.51	8.12	4.18	5.80	5.80	4.18	3.25	6.03	4.64	6.73	3.25
il	5.32	5.17	6.23	5.17	5.78	6.08	4.41	3.95	5.32	6.23	5.93	5.17	1.06
hm	—	—	—	—	—	—	—	—	—	—	—	—	0.80
ap	2.69	2.35	2.02	1.34	0.67	1.34	0.67	0.34	0.67	1.68	0.67	2.35	0.34

1. Melilite-nepheline basalt; average of 6 analyses.
2. Nepheline basalt; average of 3 analyses.
3. Nepheline basalt, ankaratrite type; only one existing analysis.
4. Picrite-basalt, mimosite type; average of 2 analyses.
5. Picrite-basalt, ankaramite type; average of 5 analyses.
6. Nepheline basanite; average of 5 analyses.
7. Analcite basanite; only one existing analysis.
8. Picrite-basalt, oceanite type; average of 9 analyses.
9. Olivine basalt; average of 53 analyses, including one olivine gabbro.
10. Andesine andesite; average of 21 analyses.
11. Basalt; average of 29 analyses.
12. Oligoclase andesite; average of 11 analyses, including one "oligoclase gabbro" (kauaiite).
13. Trachyte; average of 4 analyses.

lization of the groundmass (Edwards, 1938; Macdonald, 1940c, p. 156).

Plagioclase is the second type of phenocryst to separate. The earliest plagioclase is generally intermediate to sodic bytownite. Early plagioclase may be partly resorbed before enclosure in jackets of later more sodic plagioclase.

The period of hypersthene crystallization is less definite, and in many olivine basalts there is no evidence of hypersthene having formed

TABLE 6.—CHEMICAL ANALYSES OF LAVAS FROM SUMMIT AND FLANK VENTS OF THE SAME ERUPTIONS, AND EARLY AND LATE LAVAS FROM A SINGLE VENT

ANALYSES

	1	2	3	4	5	6	7
SiO_2	47.25	50.60	50.43	51.98	52.02	51.55	51.82
Al_2O_3	9.07	14.63	12.51	13.60	13.54	13.59	13.66
Fe_2O_3	1.45	2.38	1.62	2.54	5.25	2.33	1.50
FeO	10.41	10.15	10.23	9.02	6.58	9.04	9.68
MgO	19.96	6.38	9.09	7.36	7.31	8.02	7.24
CaO	7.88	9.56	11.10	10.26	10.23	10.31	10.09
Na_2O	1.38	2.51	1.76	2.26	2.30	2.43	2.30
K_2O	0.35	0.54	0.43	0.39	0.41	0.27	0.30
H_2O+	0.04	0.13	0.00	0.04	0.06	0.07	0.11
H_2O-	0.08	—	0.09	—	—	0.02	0.01
TiO_2	1.61	2.29	2.22	2.09	2.09	1.98	2.07
P_2O_5	0.21	0.43	0.29	0.26	0.27	0.24	0.39
MnO	0.13	0.12	0.12	0.08	0.09	0.12	0.13
Cr_2O_3	0.12	0.09	0.11	0.03	0.04	0.08	0.04
BaO	—	none	—	none	none	—	tr.
SrO	—	none	—	none	none	—	—
NiO	0.09	—	—	—	—	—	—
SO_3	—	0.07	—	0.03	0.03	—	—
CO_2	—	none	—	none	none	—	—
Total	100.03	99.88	100.00	99.94	100.22	100.05	99.36

NORMS

Q	—	3.48	2.10	5.28	8.40	2.82	4.68
Or	2.22	2.78	2.22	2.22	2.22	1.67	1.67
Ab	11.53	20.96	14.67	18.86	19.39	20.44	19.39
An	17.51	27.24	25.30	25.85	25.30	25.30	26.13
Di	16.65	14.38	22.62	19.19	18.31	20.30	17.28
Hy	18.12	21.64	25.54	20.02	13.97	21.72	22.97
Ol	28.17	—	—	—	—	—	—
Mt	2.09	3.48	2.32	3.71	7.66	3.25	2.09
Il	3.04	4.41	4.26	3.95	3.95	3.80	3.95
Ap	0.34	1.01	0.67	0.67	0.67	0.34	1.01

1. Picrite-basalt, lava of the lower vents of the 1840 eruption of Kilauea, at Nanawale Bay. G. Steiger, analyst. Cross (1915, p. 44); norm from Washington (1923c, p. 347).

2. Basalt, lava of the upper vents of the 1840 eruption of Kilauea, northwest of Makaopuhi Crater and west of Kane Nui o Hamo. F. A. Gonyer, analyst.

3. Groundmass of 1840 lava at Nanawale Bay; derived by subtracting 29 per cent of olivine of composition determined by Aurousseau and Merwin (1928, p. 560) from column 1 of this table, and recalculating to 100 per cent (Macdonald, 1944a, p. 185).

4. Basalt, latest lava of the 9200-foot vent of 1942 eruption of Mauna Loa, from aa channel 150 feet east of the main cone. F. A. Gonyer, analyst.

5. Basalt, earliest lava of the 9200-foot vent of the 1942 eruption of Mauna Loa, along the eruptive fissure 0.4 mile west of the main cone. F. A. Gonyer, analyst.

6. Basalt, lava of the flank vent of the 1926 eruption of Mauna Loa, collected at 1440 feet altitude near the highway in Kona. E. S. Shepherd, analyst. Shepherd (1938, p. 336).

7. Basalt, pumiceous phase of the lava of the 1926 eruption of Mauna Loa, near the summit of the mountain. Total includes 0.02 per cent S.; E. S. Shepherd, analyst. Shepherd (1938, p. 335).

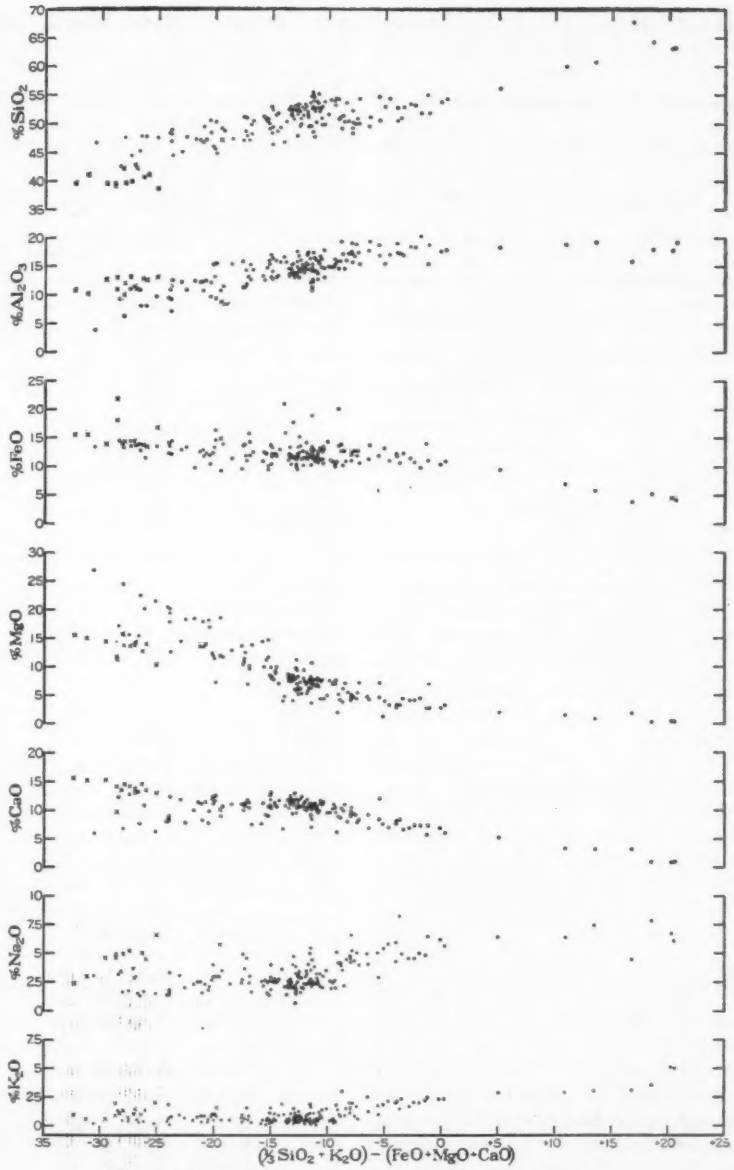

FIGURE 8.—VARIATION DIAGRAM OF HAWAIIAN ROCKS

Rocks containing modal nepheline are represented by crosses; other rocks by dots.

at any stage. However, when hypersthene is present, it commonly forms small phenocrysts, many of them enclosed in shells of monoclinic pyroxene. Its crystallization may extend well into the groundmass period, forming poikilitic grains which enclose smaller grains of the other groundmass constituents.

The monoclinic pyroxene formed during in-

tratelluric crystallization is augite. In some rocks zoning is detectable in the augite phenocrysts and indicates a gradual increase in iron content. In a small proportion of rocks, the groundmass pyroxene also is augite, but in most it is pigeonite. The latter is probably

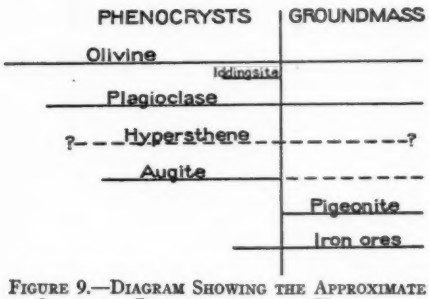

FIGURE 9.—DIAGRAM SHOWING THE APPROXIMATE ORDER OF CRYSTALLIZATION IN HAWAIIAN OLIVINE BASALTS

metastable and in at least one instance appears to have broken down to yield a mixture of augite and hypersthene (Macdonald, 1944b).

In a few rocks magnetite occurs as small phenocrysts, probably formed late in the intratelluric period, but in most the crystallization of magnetite and ilmenite is confined to the groundmass.

Crystallization Differentiation

Evidences of crystal settling.—The concentration of mafic phenocrysts in the lower part of lava flows at Kilauea, leaving the upper part relatively free of phenocrysts, was early noted by Clarence King (1878). Since then, many examples of sinking of olivine phenocrysts in flows have been observed. As would be expected, augite phenocrysts also accumulate in the lower portions of some flows, leaving the upper part nearly or quite devoid of phenocrysts. Evidence of sinking of the lighter mineral feldspar is much less abundant, but at least two good examples have been described. Basalt pahoehoe exposed along the Kolekole Pass road in the Waianae Range of Oahu consists of flow units 2–4 feet thick, in which the upper 6–10 inches is poor in phenocrysts, but grades down into a basal portion containing abundant phenocrysts of calcic plagioclase and augite (Stearns, 1940, p. 47). Similar flow units of feldspar-rich olivine basalt pahoehoe crop out along the trail on the eastern wall of Waipio Valley, in the Kohala Mountains of Hawaii. The upper 2–8 inches of each flow unit is nearly barren of phenocrysts, but grades downward into rock containing 30–50 per cent of plagioclase phenocrysts up to a centimeter long. The phenocrysts are zoned from medium bytownite to sodic labradorite (Macdonald, 1949).

Thus in the ordinary basaltic magma of Hawaii, in the condition in which it reaches the surface, crystals of all the minerals known to form in that magma as phenocrysts are denser than the magma and will sink if they are not prevented from doing so by turbulence or viscosity of the magma.

In some instances, there appears to be good evidence of the sinking of phenocrysts in upper levels of the magma columns. During the Kilauea eruption of 1840, lavas erupted high on the flanks of the volcano contained only rare small phenocrysts of olivine, whereas those erupted lower on the flank and a little later, and presumably derived from a lower level in the magma column, contained very abundant large phenocrysts of olivine (Macdonald, 1944a). The chemical compositions of the lavas from the lower and upper vents are shown respectively in columns 1 and 2 of table 6. In column 3 the observed proportion of olivine phenocrysts has been subtracted from the lava of the lower vent, and the difference recalculated to 100 per cent. The close similarity between columns 2 and 3 demonstrates that the difference in composition of the lavas of the lower and upper vents results largely from the abundant olivine phenocrysts in the lava of the lower vent.

On the other hand, analyses of the lavas of the summit and flank flows of the 1926 eruption of Mauna Loa (Table 6, columns 6 and 7) are closely similar, giving no evidence of difference in the abundance of phenocrysts at the levels in the magma column tapped by those flows. Also, analyses of the earliest and latest lavas erupted at the flank vent at 9200 feet altitude during the 1942 eruption of Mauna Loa (Table 6, columns 4 and 5) are essentially identical, except for the stage of oxidation of the iron, and give no evidence of an increase in olivine phenocrysts as lower levels in the magma column were drained during that eruption.

It is possible, of course, that the lavas of the flank eruptions during 1926 and 1942 were derived entirely from upper parts of the magma column, and thus give no indication of the composition of the magma at lower levels.

Origin of dunite inclusions.—The inclusions of dunite which occur in many Hawaiian lavas, from nepheline basalt to oligoclase andesite, have generally been attributed to the accumulation of sunken olivine crystals (Powers, 1920; Macdonald, 1942b, p. 320). This theory of origin still appears to be essentially correct. The texture of the dunite indicates, however, that simple accumulation of sunken crystals cannot be the whole explanation. The olivine grains in the dunite do not exhibit the euhedral or the rounded and embayed outlines typical of phenocrysts in the olivine basalts and picrite-basalts. Instead, almost all the grains are anhedral. Moreover, the average grain size commonly is only about 1.5 mm, in contrast to the typically much larger phenocrysts in the olivine basalt and picrite-basalt. Clearly, if the rock originated by accumulation of sunken olivine phenocrysts, as appears likely, the mass of olivine must have been recrystallized. Such recrystallization does not necessarily involve remelting, but may take place in the solid state under the influence of high temperature and pressure. Indeed, the texture is reminiscent of certain hornfels textures developed during thermal metamorphism. Some of the reduction in grain size may result from crushing.

Once the mass of olivine crystals has accumulated, and recrystallized to form the dunite rock in its familiar form, lavas rising through fissures cutting the dunite tear loose fragments of it and carry them upward as xenoliths. Many of the dunite fragments in the lava flows were clearly undergoing mechanical disintegration at the time of solidification of the flow.

Origin of rock types more salic than olivine basalt.—The basalts differ from the olivine basalts only in the dearth or absence of olivine phenocrysts and are probably derived from the olivine basalts simply by the removal of olivine phenocrysts, which sink to lower levels in the magma column or reservoir. Such rocks may become slightly oversaturated with silica (the excess appearing in the norm as quartz) by the separation and removal of olivine in more than its stoichiometric proportion.

The composition of the smallest amounts of material which must be removed from the average olivine basalt to yield the average andesine andesite, oligoclase andesite, and trachyte of the Hawaiian province is shown in Table 7. Only the principal oxides have been used in the calculation, and all iron present has been calculated as ferrous oxide. The norms of the material consist of calcic plagioclase, diopside, hypersthene, and olivine, all of which might reasonably be expected to form in basaltic magma crystallizing at shallow depth. All except hypersthene are common as phenocrysts in Hawaiian basalts extruded at the surface, although the monoclinic pyroxene actually present is an aluminous augite rather than diopside. Hypersthene phenocrysts occur only rarely in

TABLE 7.—COMPOSITION OF SMALLEST AMOUNT OF MATERIAL WHICH SUBTRACTED FROM AVERAGE HAWAIIAN OLIVINE BASALT YIELDS HAWAIIAN ROCKS MORE SALIC THAN OLIVINE BASALT

	1	2	3
Analyses			
SiO_2	50.0	49.1	48.1
Al_2O_3	12.3	12.5	13.3
Total iron as FeO	11.7	12.0	13.0
MgO	12.8	12.6	11.5
CaO	12.0	12.5	12.3
Na_2O	1.2	1.3	1.8
K_2O	0.0	0.0	0.0
Amount removed (%)	64	70	86
Norms			
ab	10.0	11.0	15.2
an	28.4	28.4	28.1
di wo	13.0	14.0	13.8
di en	7.4	7.9	7.3
di fs	5.0	5.5	6.1
hy en	13.9	8.5	1.6
hy fs	9.5	6.1	1.3
ol fo	7.5	10.6	13.9
ol fa	5.4	8.1	12.7
Mol. ratio MgO/FeO	1.96	1.89	1.59
Composition of plagioclase	An_{74}	An_{72}	An_{65}

1. Material subtracted from average olivine basalt to yield average andesine andesite.
2. Material subtracted from average olivine basalt to yield average oligoclase andesite.
3. Material subtracted from average olivine basalt to yield average trachyte.

Hawaiian lavas and are small. However, it is conceivable that hypersthene may form at depth under high pressures, becoming unstable and being rapidly resorbed as the magma rises to regions of lesser pressure (Larsen, 1940, p. 925).

TABLE 8.—COMPOSITION OF SMALLEST AMOUNT OF MATERIAL WHICH ADDED TO AVERAGE HAWAIIAN OLIVINE BASALT YIELDS HAWAIIAN ROCKS LESS SALIC THAN OLIVINE BASALT

	1	2	3	4	5	6
SiO_2	44.5	29.5	17.7	0.0	0.0	0.0
Al_2O_3	0.0	6.3	0.0	9.2	1.4	4.2
Iron as FeO	13.7	18.3	24.3	25.9	23.0	26.5
MgO	38.0	31.6	32.3	2.2	33.2	27.3
CaO	3.0	13.7	17.6	9.2	27.0	29.3
Na_2O	0.7	0.0	6.9	41.8	11.7	10.3
K_2O	0.1	0.6	1.0	11.6	3.7	2.4
Amount added (%)	40	24	21	6	19	22

1. Material added to olivine basalt to yield average picrite-basalt of oceanite type.
2. Material added to olivine basalt to yield average picrite-basalt of ankaramite type.
3. Material added to olivine basalt to yield average picrite-basalt of mimosite type.
4. Material added to olivine basalt to yield average nepheline basanite.
5. Material added to olivine basalt to yield average nepheline basalt.
6. Material added to olivine basalt to yield average melilite-nepheline basalt.

The proportion of crystallization represented by the removal of the material is 64 per cent to yield andesine andesite, increasing to 70 per cent and 86 per cent to yield oligoclase andesite and trachyte, respectively. This also appears reasonable. The average composition of the feldspar changes from An_{74} in the material subtracted to yield andesine andesite, to An_{48} in the material subtracted to yield trachyte. The feldspar compositions correspond well with those found in actual phenocrysts, and the increase in soda in the feldspar should be expected as crystallization progresses.

Similar calculations have been made for the average andesites and trachyte of the island of Hawaii, using as the hypothetical parent magma the average lava of Kilauea (Macdonald, 1949). The results differ little from those obtained here.

It thus appears possible to derive the Hawaiian andesites and trachytes from olivine basalt by simple crystallization differentiation.

Origin of picrite-basalts of oceanite type.—The picrite-basalts rich in large phenocrysts of olivine, termed chrysophyric basalts by J. D. Dana, have very generally been attributed to the accumulation of olivine phenocrysts in basaltic magma (Bowen, 1927; 1928, p. 159–164; Daly, 1911; 1933, p. 396–397; Shand, 1943, p. 321). Table 8, column 1, shows the calculated composition of the smallest amount of material which added to the average Hawaiian olivine basalt would yield a rock of the composition of the average picrite-basalt of the oceanite type. The material corresponds to a mixture of about 70 per cent olivine (Fo_{77}), 17 per cent hypersthene, 12 per cent diopside, and negligible amounts of sodium and potassium orthosilicate. This essentially confirms the hypothesis, supported also by microscopic study, that the picrite-basalts of oceanite type differ little from the olivine basalt except in the abundance of olivine phenocrysts. They are unquestionably derived by the local enrichment of olivine basalt magma in crystals of olivine, and possibly small amounts of pyroxene, settled from higher levels in the magma body.

Daly (1933, p. 398) has called attention to the greater amount of iron in the picrite-basalts of oceanite type than in the Hawaiian olivine basalts and has suggested that some iron oxide may have accompanied the olivine in its concentration. The material of column 1 in Table 8 contains a slightly larger proportion of iron than the olivine phenocrysts occurring in the rocks. The iron oxide was probably added as small crystals of magnetite enclosed in the olivine phenocrysts.

Daly (1933, p. 396) believes that the rather uniform dispersal of the olivine phenocrysts in the oceanite type of picrite-basalt is evidence against the idea that they represent merely accumulated early-formed crystals. He believes that the sunken crystals were dissolved in the superheated magma and later recrystallized, thus explaining their occurrence as uniformly distributed isolated crystals rather than as clots and rafts of crystals. Bowen (1928, p. 164), however, believes that olivine in excess of 12 or 15 per cent in basaltic rocks represents an excess of crystalline material which was never in solution in the enclosing magma. There is little

direct evidence in Hawaiian rocks bearing on this difference of opinion. The olivine phenocrysts have quite generally been partly resorbed, as would be expected if they settled into deeper regions of hotter magma still undersaturated in olivine. But the resorption appears to have been small in amount. There is nothing suggesting complete or even extensive resolution. Moreover, there appears to be no very conclusive reason why olivine phenocrysts settling freely in a magma mildly stirred by thermal and multiphase convection should become aggregated into clots. The dunite inclusions mentioned earlier may represent such phenocrysts the sinking of which was interrupted by some obstacle, resulting in the formation of nearly monomineralic masses. The lack of any nonporphyritic rocks containing more than 12 or 15 per cent of normative olivine, pointed out by Bowen, holds true in Hawaii; and it seems unlikely that if such magmas existed they would fail to appear at the surface, under the conditions of free rapid eruption which characterize Hawaiian volcanoes during most of their existence. The evidence definitely favors Bowen's conclusion, that much of the olivine in the picrite-basalts of oceanite type was never in solution in the enclosing magma.

Origin of the picrite-basalts of ankaramite type.—The picrite-basalts of ankaramite type also have generally been considered to result from the accumulation in basaltic magma of mafic phenocrysts (Washington and Keyes, 1928, p. 219; Bowen, 1928, p. 165; Daly, 1933, p. 398; Shand, 1943, p. 349; Macdonald, 1942b, p. 310). The present study does not entirely confirm this assumption. If such were the case, the groundmass should consist essentially of olivine basalt, modified only by minor amounts of resorption of the phenocrysts. Columns 1 and 2 of Table 9 show the composition of the groundmass of two examples of these lavas, determined by subtracting from the bulk composition of the rock the composition of the augite and olivine phenocrysts, the proportions of the latter having been determined by Rosiwal thin-section analyses. The composition used for the augite phenocrysts is that determined by actual chemical analysis (Washington and Keyes, 1928, p. 213); the composition used for the olivine phenocrysts is that determined for phenocrysts in the 1840 lava of Kilauea (Aurousseau and Merwin, 1928), and approximately confirmed by the optical properties of the olivine actually present. It is immediately obvious that the groundmass of the two lavas differs markedly in composition from olivine basalt. The rocks cannot be the result of simple addition, without resolution, of phenocrysts of olivine and augite.

The possibility must be considered that the rock is the result of enrichment in sunken crystals, some of which have been redissolved to modify the composition of the groundmass. If this is the case, the bulk composition of the rock should equal olivine basalt plus the added sunken crystals, recalculated to 100 per cent. Column 2 of Table 8 shows the calculated composition of the smallest amount of material which added to the average Hawaiian olivine basalt yields the bulk composition of the average picrite-basalt of the ankaramite type. Obviously the material is not of a sort which would be expected to separate as crystals during the early stages of normal crystallization of a basaltic magma. Doubling or even trebling the amount of material added, to produce the same result, of course alters the composition of the added material in the direction of olivine basalt, but the material is still not such as would be expected to crystallize from olivine basalt. Column 4 of Table 9 shows the composition of the smallest amount of material which would have to be assimilated by the average olivine basalt magma to yield the average groundmass of the picrite-basalts of ankaramite type (Table 9, column 3). Column 5 shows the composition of material yielding a similar product, but assuming more material assimilated. Obviously again, the material of neither column is such as would be derived by normal crystallization of olivine basalt, and the proportion of added material (55 per cent of the resultant mixture, or more than its own volume) assumed to have been dissolved by the magma in column 5 is wholly unreasonable. It is concluded that the ankaramite type of picrite-basalt cannot be derived by simple enrichment of olivine basalt magma with crystals formed in another part of the magma body.

Another possibility is to consider the picrite-basalts of ankaramite type as residual from the crystallization of olivine basalt, just as are the andesites and trachytes. The composition of the smallest amount of material which removed

TABLE 9.—TOTAL AND GROUNDMASS COMPOSITIONS OF PICRITE-BASALT OF ANKARAMITE TYPE, AND MATERIAL ADDED TO OLIVINE BASALT TO PRODUCE THEM

	1	2	3	4	5	6	7
SiO_2	44.1	45.3	44.7	23.2	40.0	45.5	45.4
Al_2O_3	16.0	13.2	14.6	17.3	15.2	11.9	11.8
Iron as FeO	17.5	15.2	16.3	33.4	20.1	13.3	13.4
MgO	5.3	10.4	7.9	0.0	6.2	15.3	15.3
CaO	13.8	13.4	13.6	24.0	15.9	11.5	11.8
Na_2O	2.6	1.8	2.2	1.1	1.9	1.9	1.9
K_2O	0.7	0.7	0.7	1.0	0.7	0.6	0.4
Amount added (%)	—	—	—	21	55	—	—

1. Groundmass of "Big-augite" flow, near western rim of Haleakala Crater, Maui. Composition derived by subtracting 21 per cent augite phenocrysts and 20.5 per cent olivine phenocrysts from the bulk composition determined by analysis (Macdonald and Powers, 1946), and recalculating to 100 per cent.

2. Groundmass of flow near base of Namanaokeakua Cone, Haleakala Crater, Maui. Composition determined by subtracting 12 per cent olivine and 15 per cent augite phenocrysts from the bulk composition determined by analysis (Cross, 1915, p. 29; Macdonald, 1942b, p. 309), and recalculating to 100 per cent.

3. Average of columns 1 and 2.

4. Smallest amount of material which added to average olivine basalt yields the average groundmass of column 3.

5. Material added to average olivine basalt to obtain the average groundmass of column 3, assuming a larger amount of added material than in column 4.

6. Average picrite-basalt of ankaramite type; principal oxides, recalculated to 100 per cent.

7. Average olivine basalt (72 per cent), plus 17 per cent olivine, 5 per cent anorthite, 3 per cent CaO, and 3 per cent FeO.

TABLE 10.—COMPOSITION OF SMALLEST AMOUNT OF MATERIAL WHICH SUBTRACTED FROM AVERAGE HAWAIIAN OLIVINE BASALT YIELDS HAWAIIAN ROCKS LESS SALIC THAN OLIVINE BASALT

ANALYSES

	1	2	3	4	5
SiO_2	60.0	64.2	53.2	59.9	66.5
Al_2O_3	17.5	19.8	14.2	16.1	16.9
Iron as FeO	8.7	6.6	11.0	9.6	6.9
MgO	0.0	2.8	10.4	5.8	4.8
CaO	9.6	8.0	10.9	7.8	4.9
Na_2O	3.6	0.6	0.3	0.8	0.0
K_2O	0.6	0.0	0.0	0.0	0.0
Material subtracted (%)	35	33	54	50	40

NORMS

Q		13.0	28.0	6.9	21.8	43.0
or		3.3	—	—	—	—
ab		30.4	13.6	2.6	6.8	—
an		30.0	39.8	37.2	38.6	24.5
c		—	2.6	—	0.6	8.0
di	wo	7.3	—	7.1	—	—
	en	—	—	3.8	—	—
	fs	8.3	—	3.0	—	—
hy	en	—	7.0	22.2	14.5	12.0
	fs	7.7	12.1	17.2	17.6	12.7
Mol. ratio MgO/FeO		0	0.76	1.70	1.09	1.25
Composition of plagioclase		An_{50}	An_{75}	An_{93}	An_{85}	An_{90}

1. Material subtracted to yield average picrite-basalt of ankaramite type.

2. Material subtracted to yield average picrite-basalt of mimosite type.

3. Material subtracted to yield average nepheline basanite.

4. Material subtracted to yield average nepheline basalt.

5. Material subtracted to yield average melilite-nepheline basalt.

from olivine basalt would leave the rocks of ankaramite type is shown in column 1 of Table 10. Again the material is not such as would be expected to separate during normal crystallization of olivine basalt. The norm of the material contains 13 per cent quartz, and the plagioclase (An_{50}) is even slightly more sodic than the average plagioclase of the parent olivine basalt (An_{54}). Increasing the degree of crystallization to about 60 per cent (Table 11, column 1) eliminates quartz from the norm, but the feldspar separated (An_{68}) is still more sodic than that of the parent olivine basalt, whereas at 60 per cent crystallization according to the well-known feldspar diagram (Bowen, 1928, p. 34) the feldspar should have a composition of about An_{86}. It is virtually certain, therefore, that the picrite-basalts of ankaramite type cannot have originated as a rest-magma resulting from the partial crystallization of olivine basalt. The process appears unlikely at

any rate, as its existence would mean the derivation of two strongly contrasted end magmas from the same parent magma, in the same chamber, at the same time, and by the separation of essentially the same sort of crystalline material.

Assuming a submagma derived through the assimilation in olivine basalt of 10 per cent of sunken olivine phenocrysts and 30 per cent of the crystalline mixture removed from olivine basalt during the formation of andesine andesite (Table 7), the picrite-basalts of ankaramite type can be derived in theory by the removal of material essentially similar to that which should form during the crystallization of basaltic magma. The composition of the smallest amount of material which, separated from such a cafemic submagma, would yield the ankaramite type of picrite-basalt is shown in column 1 of Table 12. It contains 6.2 per cent normative quartz. If the degree of crystallization is increased sufficiently to eliminate quartz from the norm, the material removed has the composition shown in column 1, of Table 13. The general nature of the material is not unlike that which would be expected to crystallize from a magma of the composition assumed. However, the normative plagioclase (An_{82}) is more sodic than would be expected, on the basis of the feldspar diagram, at the degree of crystallization indicated (55 per cent). With normative feldspar of composition An_{48} in the parent magma, at 50 per cent crystallization the plagioclase separating should have a composition of about An_{75}, and if reaction between the solid and liquid phases had not been complete the average of the feldspar separated should be even more calcic. If this were the only objection to the process it might perhaps be ignored, because of inaccuracy of the data and the necessarily rough nature of the calculations. However, a still greater objection is that the large amount of resolution and recrystallization involved is inconsistent with the fairly early appearance of the rocks in question in the volcanic cycle and their fairly great abundance.

Thus, the picrite-basalts of ankaramite type probably are not the result of pure crystallization differentiation. Some influence outside the igneous body, operating in addition to crystallization, may therefore be sought. It is readily found in the addition of small amounts of lime-

TABLE 11.—COMPOSITIONS OF MATERIAL WHICH SUBTRACTED FROM AVERAGE OLIVINE BASALT YIELDS HAWAIIAN ROCK TYPES LESS SALIC THAN OLIVINE BASALT

Assuming a greater degree of crystallization of the parent magma than in Table 10

	1	2	3	4	5	6
ANALYSES						
SiO_2	53.8	51.9	53.0	53.1	53.4	50.5
Al_2O_3	15.2	14.0	15.0	14.4	14.4	13.8
Iron as FeO	10.6	11.3	10.7	11.1	10.8	11.7
MgO	6.6	10.2	8.4	9.0	9.1	10.1
CaO	10.3	10.9	10.3	9.9	9.7	10.8
Na_2O	2.9	1.4	2.1	2.1	2.1	2.5
K_2O	0.6	0.3	0.5	0.4	0.5	0.6
Amount removed (%)	61	70	73	78	79	—
NORMS						
or	3.3	1.7	2.8	2.2	2.8	3.3
ab	24.6	12.0	17.8	17.8	17.8	21.0
an	26.7	30.9	30.0	29.2	28.4	25.0
di wo	10.2	9.7	8.8	8.4	8.2	12.0
di en	4.7	5.2	4.5	4.3	4.3	6.2
di fs	5.4	4.2	4.1	3.8	3.7	5.4
hy en	10.5	19.2	16.5	18.1	18.5	3.0
hy fs	12.3	15.6	15.6	16.4	16.1	2.6
ol fo	0.9	0.8	—	—	—	11.1
ol fa	1.3	0.7	—	—	—	10.4
Mol. ratio MgO/FeO	1.12	1.62	1.41	1.46	1.52	1.53
Plagioclase	An_{82}	An_{75}	An_{83}	An_{83}	An_{81}	An_{84}

1. Material removed from olivine basalt to yield average picrite-basalt of ankaramite type.

2. Material removed from olivine basalt to yield average nepheline basanite.

3. Material removed from olivine basalt to yield average picrite-basalt of mimosite type.

4. Material removed from olivine basalt to yield average nepheline basalt.

5. Material removed from olivine basalt to yield average melilite-nepheline basalt.

6. Average olivine basalt, principal oxides recalculated to 100 per cent.

stone to the olivine basalt. This process has been suggested by Daly (1944) to explain the origin of the Hawaiian nepheline-bearing rocks, but has not previously been applied to the picrite-basalts of ankaramite type. Besides lime, it is necessary to add crystals of olivine and calcic plagioclase, minerals known to form phenocrysts early in the crystallization of olivine basalt and

TABLE 12.—COMPOSITIONS OF CAFEMIC SUBMAGMA AND OF SMALLEST AMOUNTS OF MATERIAL WHICH SUBTRACTED FROM IT YIELD HAWAIIAN ROCKS LESS SALIC THAN OLIVINE BASALT

ANALYSES

	1	2	3	4	5	6
SiO_2	56.8	53.9	50.3	53.4	55.7	49.3
Al_2O_3	12.6	13.1	11.4	12.4	12.4	12.1
Iron as FeO	8.4	9.6	11.3	10.6	9.5	11.7
MgO	13.0	14.3	16.6	14.6	15.0	14.5
CaO	7.4	8.4	9.8	8.2	7.0	10.1
Na_2O	1.8	0.7	0.6	0.8	0.4	1.9
K_2O	0.0	0.0	0.0	0.0	0.0	0.4
Amount removed (%)	33	56	70	67	60	—

NORMS

Q		6.2	10.5	—	2.7	8.6	—
or		—	—	—	—	—	2.2
ab		15.2	5.8	5.2	6.8	3.1	16.2
an		26.4	32.5	28.4	30.3	32.2	23.4
di	wo	4.3	3.8	8.5	4.3	1.0	11.1
	en	2.7	2.4	5.3	2.6	0.6	6.6
	fs	1.3	1.2	2.6	1.4	0.4	4.0
hy	en	29.8	23.3	26.0	33.9	36.9	5.2
	fs	14.1	16.4	12.8	18.0	17.0	3.0
ol	fo	—	—	7.1	—	—	17.1
	fa	—	—	4.1	—	—	11.2
Mol. ratio MgO/FeO		2.78	2.68	2.64	2.48	2.84	2.22
Plagioclase		An_{62}	An_{85}	An_{85}	An_{87}	An_{91}	An_{59}

1. Material subtracted from the cafemic submagma (column 6) to yield average picrite-basalt of the ankaramite type.

2. Material subtracted to yield average picrite-basalt of the mimosite type.

3. Material subtracted to yield average nepheline basanite.

4. Material subtracted to yield average nepheline basalt.

5. Material subtracted to yield average melilite-nepheline basalt.

6. Cafemic submagma, composed of 60 per cent average olivine basalt, 30 per cent material of the composition of that removed from olivine basalt in the formation of andesine andesite, and 10 per cent olivine.

which may logically be expected to sink into lower levels of the magma. Assuming the addition to average olivine basalt of 17 per cent of olivine (approximately the average excess of olivine phenocrysts in the rocks of ankaramite type over the amount of normative olivine in the parent olivine basalt), 5 per cent anorthite, 3 per cent CaO, and 3 per cent FeO, the result shown in column 7 of Table 9 is obtained. The actual plagioclase added by crystal settling would be more sodic, but anorthite is assumed for simplicity in calculation. Addition of a slightly greater amount of more sodic plagioclase would produce essentially the same result. The correspondence in composition to the average picrite-basalt of ankaramite type (Table 9, column 6) is very close. Inasmuch as the volatile constituents must be ignored in such calculations, the CO_2 of the limestone does not appear in the calculations. Its presence, however, may contribute to the obvious gas-richness of the eruptions of these lavas, which appear to have been just as explosive as those of the associated andesites and much more so than those of the ordinary olivine basalts. This richness in gas would be less expectable in a rock formed entirely by the concentration in olivine basalt magma of a gas-free sunken crystal phase, presumably reducing rather than augmenting the proportion of gas in the magma. The iron oxide may be added principally as magnetite crystals, many of which occur as inclusions in the olivine phenocrysts, or it may be concentrated by volatile transfer, as has been shown to take place in certain Keweenawan lavas of the Lake Superior district (Broderick, 1935).

The picrite-basalts of ankaramite type probably have resulted from enrichment of olivine basaltic magma in sunken crystals of olivine, calcic plagioclase, and magnetite from higher levels in the magma body, together with the assimilation of a small amount of limestone.

Origin of nepheline-bearing rocks and picrite-basalt of mimosite type.—The Hawaiian nepheline basanites, nepheline basalts, melilite-nepheline basalts, and picrite-basalts of the mimosite type form a group closely related in time and space, and presumably also in origin. The general features of the group which must be explained in any theory of petrogenesis are: low silica content; rather high content of magnesia and iron oxide; high lime content; high content of alkalies, particularly soda; high volatile content of the magma, indicated by the large amount of lava fountaining which accompanied many of the eruptions, and by the presence of pegmatitoid veinlets in some of the

TABLE 13.—COMPOSITIONS OF MATERIAL SUBTRACTED FROM THE CAFEMIC SUBMAGMA TO YIELD HAWAIIAN ROCKS LESS SALIC THAN OLIVINE BASALT

Allowing a greater degree of crystallization than in Table 12.

ANALYSES

	1	2		3		4		5	
SiO_2	52.4	51.2	49.9	49.9	49.5	52.1	50.2	52.5	50.4
Al_2O_3	12.3	12.6	12.3	11.6	11.9	12.3	12.2	12.2	12.2
Iron as FeO	10.4	10.8	11.4	11.4	11.6	11.0	11.5	10.7	11.3
MgO	13.8	14.4	14.5	16.2	15.1	14.5	14.5	14.8	14.6
CaO	8.9	9.4	9.9	9.9	10.0	8.9	9.7	8.5	9.6
Na_2O	1.9	1.4	1.7	0.9	1.6	1.1	1.6	1.1	1.6
K_2O	0.3	0.2	0.3	0.1	0.3	0.1	0.3	0.2	0.3
Amount removed (%)	55	75	90	75	90	75	90	75	90

NORMS

or	1.7	1.1	1.7	0.6	1.7	0.6	1.7	1.1	1.7
ab	16.2	12.0	14.2	7.9	13.6	9.4	13.6	9.4	13.6
an	24.2	27.5	25.3	27.2	24.5	28.4	25.3	27.8	25.0
di wo	8.4	8.0	9.8	9.2	10.6	6.6	9.5	6.0	9.4
di en	5.1	4.9	6.0	5.7	6.4	4.0	5.7	3.7	5.7
di fs	2.8	2.6	3.4	2.9	3.6	2.2	3.3	2.0	3.2
hy en	20.2	20.8	11.4	20.2	9.7	29.0	13.4	30.9	12.8
hy fs	11.0	11.2	6.3	10.3	5.5	16.4	7.9	16.6	7.3
ol fo	6.4	7.2	13.2	10.2	15.1	2.2	12.0	1.7	12.6
ol fa	4.1	4.6	8.6	5.9	9.4	1.2	7.6	0.8	8.0
Mol. ratio MgO/FeO	2.40	2.40	2.29	2.56	2.34	2.37	2.26	2.48	2.32
Plagioclase	An_{60}	An_{70}	An_{64}	An_{78}	An_{64}	An_{75}	An_{68}	An_{75}	An_{68}

1. Material subtracted to yield average picrite-basalt of the ankaramite type, assuming 55 per cent crystallization of the cafemic submagma.

2. Material subtracted to yield average picrite-basalt of the mimosite type, assuming respectively 75 and 90 per cent crystallization.

3. Material subtracted to yield average nepheline basanite, assuming respectively 75 and 90 per cent crystallization.

4. Material subtracted to yield average nepheline basalt, assuming respectively 75 and 90 per cent crystallization.

5. Material subtracted to yield average melilite-nepheline basalt, assuming respectively 75 and 90 per cent crystallization.

flows; and finally their appearance only after long periods of volcanic quiescence.

The low silica content suggests desilication of the magma. The high content of magnesia, lime, and iron oxide suggests that the magma has been made more basic. The high volatile content suggests either the addition of extraneous volatiles or the concentration of juvenile volatiles into a relatively small end fraction of the magma. The appearance of the rocks only after long quiescence suggests the necessity for long-continued quiet crystallization in the magma reservoir.

Desilication and an increase in lime and volatiles could be accounted for by the assimilation of calcareous sediments, as suggested by Daly (1944). Desilication and increase in lime, magnesia, and iron also could result from the accumulation of sunken early-formed crystals of olivine and calcic plagioclase. But the mere addition to olivine basalt magma of limestone, or olivine and calcic plagioclase, or both, will not account for the high alkali content, nor will it explain the necessity of a long quiet period preceding the eruption of the rocks. Both of the latter features can, however, be explained

if the rocks are end magmas resulting from long crystallization of a cafemic submagma, enriched in sunken crystals and possibly in sedimentary lime. Even without the addition of limestone, such an end magma would be rich in volatiles.

Table 8 shows the compositions of the material which would have to be added to average olivine basalt to produce the nepheline-bearing rocks and the mimosite type of picrite-basalt. Obviously such material would be most unlikely to crystallize from olivine basalt. Table 10, columns 2 to 5, shows the composition of the smallest amount of material the removal of which from olivine basalt would yield the rocks in question. Again, the material is not such as would be expected to crystallize from olivine basalt magma. The norms all contain a notable amount of quartz, which is very unlikely to form during the early or intermediate stages of crystallization, even though it might form during late stages owing to earlier excessive crystallization of olivine. Moreover, in columns 4 and 5 the plagioclase is considerably more calcic than would be expected as the average of the plagioclase separated from olivine basalt at 50 per cent crystallization. Column 2 contains no magnesia, column 3 contains very little, and all contain notably less magnesia than iron oxide. That condition is not normal for material separated during early stages of crystallization of olivine basalt, in which magnesia nearly always greatly exceeds iron oxide.

These compositions represent, however, the smallest amounts of material which could be subtracted from olivine basalt to yield the less salic rock types. The possibility of a greater degree of crystallization of the parent magma must be considered. In Table 11, a sufficiently greater degree of crystallization has been assumed to remove the excess of silica, expressed as quartz in the norm. The material approaches much more closely in composition that which would be expected to form as crystals in an olivine basalt magma, but important minor discrepancies still remain. During the normal course of crystallization of olivine basalt the early-formed femic minerals are rich in magnesia, and the proportion of iron relative to magnesia increases with the degree of crystallization. At no stage should the plagioclase of the crystalline phase be richer in soda than the parent magma, nor the femic minerals of the crystalline phase be richer in iron than the parent magma. The parent olivine basalt (Table 11, column 6) has a molecular ratio MgO/FeO of 1.53. In the material of column 3 the MgO/FeO ratio is decidedly lower than in the olivine basalt, indicating that this material is unlikely to have separated during the crystallization of the olivine basalt. In columns 4 and 5 the ratio is only slightly lower than in the olivine basalt, but, considering the fact that the material represents crystallization of only 80 per cent of the parent magma, it again appears unlikely that such material would form during normal crystallization of olivine basalt magma. Some other mode of origin should therefore be sought for the picrite-basalts of mimosite type and the nepheline basalts. Column 2 represents the material the removal of which from olivine basalt would yield nepheline basanite. It appears possible that such material might crystallize during the solidification of olivine basalt magma, and hence that nepheline basanite might originate as a rest-magma resulting from the fractional crystallization of olivine basalt. However, the occurrence of the basanites is so closely related to that of the nepheline basalts that it is very unlikely that the modes of origin of the two rocks differ in more than minor detail and degree.

Although neither simple addition nor simple subtraction of material from the parent olivine basalt appears adequate to account for the nepheline basalts and picrite-basalts of the mimosite type, there remains to be tried a combination of the two processes. The olivine removed from olivine basalt during the formation of basalt, and the material of more complex composition believed to have been removed by crystal differentiation during the formation of the andesites and trachytes (Table 7) sink to lower levels in the magma. There they encounter higher temperatures, and it is reasonable to suppose that part or all of them are resorbed, altering the composition of the liquid magma. This submagma, richer in lime, iron oxide, and magnesia than the original olivine basalt, may be supposed in time also to reach the temperature at which crystallization commences, and fractional crystallization, with the sinking of crystals, to result in the formation of a rest-magma still different in composition from

any magma previously formed during the magmatic cycle.

In Tables 12 and 13, there is assumed a submagma derived by the addition to olivine basalt of 10 per cent olivine, of the composition of that forming phenocrysts in the 1840 lava flow of Kilauea, and 30 per cent of the crystalline mixture removed from olivine basalt during the formation of andesine andesite (Table 7). The writer has calculated the composition of material which would have to be subtracted from such a cafemic submagma to yield the picrite-basalts and nepheline-bearing lavas. In Table 12 the smallest possible amounts of subtracted material are shown, whereas in Table 13 a more advanced crystallization of the submagma is assumed. The smallest possible amount of material which subtracted from the submagma would yield the nepheline basanite is shown in column 3 of Table 12. Its composition appears entirely possible. However, the smallest possible amounts of material subtracted to form the nepheline and melilite-nepheline basalts and picrite-basalts of the mimosite type (Table 12, columns 2, 4, and 5) contain normative quartz and very calcic plagioclase and are considered unlikely. The material separated on crystallization of a larger proportion of the magma (Table 13) appears however, reasonable in composition. As compared with that removed from olivine basalt during the formation of the andesites and trachyte (Table 7), the material removed from the submagma during the formation of the nepheline-bearing lavas contains more calcic plagioclase, a larger proportion of femic minerals, and a higher ratio of MgO/FeO. This is reasonable in view of the more cafemic composition of the submagma, as compared with olivine basalt. Assumption of moderately different proportions of added material in forming the submagma alters the details of the process, but not its general character. Assumption of a still greater degree of crystallization of the submagma somewhat alters the composition of the crystallized material, but likewise does not alter its general character. Table 13 indicates the composition of the material removed under assumptions of 75 and 90 per cent crystallization. The occurrence of the nepheline-bearing lavas only after periods of erosion long enough to permit the cutting of canyons several thousand feet deep suggests that the degree of crystallization in the magma reservoirs from which they were erupted may be very high, perhaps in the vicinity of 90 per cent.

The materials removed from the cafemic submagma in deriving the mimosite type of picrite-basalt and the three types of nepheline-bearing lavas are in each case quite similar in composition at the same degree of crystallization, and the similarity increases as the degree of crystallization increases. Apparently the formation of one of these types rather than the other depends on relatively minor differences in the separated material, owing perhaps to small differences in temperature, pressure, and speed of cooling, and probably also to small differences in composition and degree of crystallization of the cafemic submagma. These rocks probably are generated in small, largely or entirely isolated residual pockets of magma, left after most of the magma body has consolidated. Among such pockets minor variations in physical conditions and magmatic composition are to be expected. Under low pressures most of the mafic material may crystallize as olivine, resulting in the formation of nepheline basanite end magma. In contrast, at higher pressures hypersthene may form, and the residual liquid may be nepheline basalt. Addition of a larger amount of lime in some magma chambers, either as sunken plagioclase or by assimilation of limestone, may lead to the local development of melilite rocks.

Two points may be raised in objection to the existence of the hypothetical cafemic submagma: (1) the large proportion of material redissolved in the olivine basalt magma to form it (a proportion equal to 67 per cent of the original assimilating magma); and (2) the fact that it does not itself ever appear at the surface. There is no known rock in the Hawaiian province having the composition of the cafemic submagma. These objections may be at least partly answered by agreeing that the cafemic submagma probably did not ever exist as a fluid magma. The figures for material added and material subtracted from the magma in the formation of the nepheline basalts and related rocks represent merely the summation of the changes involved. All the added material need not have been simultaneously dissolved in the magma. Heat derived by crystallization of some of the

subtracted material may have been available to aid in the solution of the added material. Furthermore, much of the added material may never have been dissolved at all, but instead may merely have reacted with the magmatic fluid to form new solid phases.

Effects of Volatile Transfer

The effects of volatile transfer in the formation of the Hawaiian rocks are difficult or impossible to evaluate. It has been demonstrated that all of the rock types can be derived from parent olivine basalt without resorting to volatile transfer. However, some enrichment in iron and alkalies may have resulted from this process, and the possible transfer of iron oxides by volatiles has been suggested in discussing the formation of the ankaramite type of picrite-basalt. Column 4 of Table 8 indicates that the material which if added to olivine basalt would yield nepheline basanite consists very largely of iron oxide and alkalies, thus resembling in composition material believed by some writers to have been added to magmas or parts of magmas by volatile transfer. Some alkalies and iron also may have been added to the nepheline basalts. It is impossible to state whether or not such volatile transfer has occurred. It can only be pointed out that the process is not necessary to explain the origin of Hawaiian lavas.

Effects of Selective Remelting

Superheated magma tends to dissolve the xenoliths and wall rocks with which it is in contact. In the Hawaiian province, the xenoliths and wall rocks encountered by the rising olivine basalt magma are very largely olivine basalt, closely similar in composition to the nonfugitive phases of the magma. If the temperature of the magma is still too high for the beginning of crystallization, the older solidified olivine basalt tends to be completely dissolved. If crystallization has already started in the magma, only the mineral phases in the solid rock which crystallized at a temperature lower than that of the magma are dissolved, although phases which crystallized at higher temperatures may be transformed by reaction with the magma. In either case, the constituents most readily redissolved are those which crystallized at the lowest temperature—the latest-crystallized components of the rock. In most of the olivine basalts the entire groundmass crystallized essentially simultaneously, but in the slower-cooled olivine gabbros the last minerals to crystallize were the alkalic feldspars, and these are undoubtedly the lowest-melting constituents of olivine basalt.

To what extent olivine basalt magma contains enough superheat to remelt older olivine basalt is debatable, and undoubtedly no single invariable answer can be given. The rounded xenoliths which are found in some Hawaiian lava flows certainly indicate at least a small amount of resorption, though the disintegration of the xenoliths is partly, and may be largely, mechanical. Moreover, crystallization of the olivine basalt magma at the time of eruption has not progressed beyond the precipitation of some of the olivine and calcic plagioclase, and the magma in the conduit should therefore be able to redissolve the low-melting sodic and potassic feldspar of the older rocks, enriching the magma in alkalies.

This process of enrichment of olivine basalt magma in alkalies, by selective resolution of the low-melting constituents of wall rocks and xenoliths, has been invoked by Daly (1933, p. 402–404) to explain the formation of such rocks as the analcite basalts, essexites, crinanites, and trachyandesites. In Hawaii, he believes that the same process, together with crystal settling in the magma, is responsible for the formation of the oligoclase andesites and trachytes (Daly, 1944, p. 1383–1384).

As Daly has pointed out (1933, p. 322), the process of selective redissolving is essentially the reverse of fractional crystallization. The effects on the magma of the redissolving of low-melting components of the wall rocks are similar to the effects of crystal differentiation, in enriching the end magma with low-melting compounds, notably sodic and potassic feldspar. The effects of the process cannot be confidently isolated from those of crystal differentiation, and consequently the process itself is difficult to evaluate. In general, however, it appears probable that most of the rock of engulfed xenoliths and assimilated wall rock—including most or all of the mineral phases crystallized at a temperature lower than that of the magma, rather than merely the lowest-melting constituents—must

be reincorporated into the liquid phase of the magma. The rounded xenoliths in Hawaiian lavas show no signs of selective resolution, apparently redissolving essentially in their entirety. Complete or nearly complete resolution of olivine basalt in olivine basalt magma should not greatly alter the bulk composition of the nonvolatile fraction of the magma. In the Hawaiian Islands, selective resolution is probably of relatively slight importance in comparison with crystal differentiation.

Theory of Limestone Assimilation

Independently of theoretical connotations, it has been concluded on an earlier page that the formation of the picrite-basalts of ankaramite type may have required the addition to the magma of a small amount of lime, probably through the assimilation of limestone. The theory of limestone assimilation has also been advanced by Daly (1944) to explain the origin of the Hawaiian nepheline basalts and especially the melilite-nepheline basalts. It has been demonstrated above that to derive these rocks it is not necessary to assimilate limestone or any other foreign substance. However, if limestone is assimilated during the formation of the picrite-basalts of ankaramite type, it is quite likely that the magma which gives rise to the nepheline-bearing rocks also has assimilated limestone. It would, moreover, furnish a ready explanation of the distinctly higher lime content of these lavas, as compared with the parent olivine basalt; and indeed it is unavoidable if the nephelinic rocks actually represent the residue of the general magma body which earlier gave rise to the other rock types, including the ankaramite type of picrite-basalt. The addition of a small amount of lime to the cafemic submagma which is believed to generate the nephelinic rocks alters the nature of the separated material only in that a greater proportion of the material goes into the formation of lime-pyroxene (normative diopside) rather than hypersthene, and the proportion of olivine is slightly increased.

Assimilation of limestone by olivine basalt magma is not in itself sufficient, even with ensuing crystal differentiation, to produce the nepheline and melilite-nepheline basalts. These rocks are ultrafemic, as well as alkali-rich and undersaturated with silica. As Daly himself points out (1933, p. 518), assimilation of lime followed by crystallization and settling out of diopside, because it removes magnesia and iron oxide along with lime, should result in a residual liquid less femic than the original magma. To derive the nepheline-bearing lavas of Hawaii it is necessary to make the magma ultrafemic by assimilation of sunken femic crystals, as well as to add lime. In fact, the addition of lime does not appear to be essential to the process.

If the assimilation of limestone is hypothecated, it must be shown that contact of the magma with limestone is possible. There is now general agreement that the coral limestones of the Hawaiian Islands are mere superficial fringes, that most of the nephelinic magmas could never have come into contact with them, and that there is no evidence of limestone interbedded with the lavas of the main shield volcanoes. Not only is there no evidence of the latter, but it is theoretically very unlikely, as the frequent lava inundations during the period of growth of the main shields would inhibit the growth of reef-forming organisms in sufficient abundance to build reefs (Winchell, 1947). Daly (1944) suggested the presence over very large areas of the ocean bottom of a layer of calcareous material 500 or 600 meters thick. The known distribution of Globigerina ooze and red clay (which also contains a large proportion of lime) lends some credence to the suggestion. Furthermore, the present paucity of calcareous deposits at great oceanic depths probably results from the fact that the cold bottom waters are undersaturated in calcium carbonate and tend to dissolve sinking calcareous tests. Because the present coldness of the bottom waters apparently is largely the result of influx of water from the frigid polar regions, in earlier geologic times when differentiation of surface temperatures of the earth was less marked, the temperature stratification of the ocean may have been much less marked than it is now. Under those conditions deposition of calcareous sediments in abyssal regions may have been much more general and abundant. The possibility of the widespread layer of limestone hypothecated by Daly cannot be denied.

The limestone layer is believed by Daly to underlie the mid-Pacific volcanoes and to be bent down by the isostatic sinking resulting

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The limestone layer is believed by Daly to underlie the mid-Pacific volcanoes and to be bent down by the isostatic sinking resulting

from the load of volcanic rock on the suboceanic crust. Contact of the limestone with the basalt magma of the ascending abyssolith results in assimilation of the limestone.

Stearns (1945) has attacked Daly's theory of limestone assimilation on the grounds that limestone on the ocean floor, buried by the volcanic pile, would not come in contact with the magma generating the nepheline basalts. Even if Stearns' arguments be accepted, they in no way preclude the assimilation of limestone during an earlier magmatic stage. Stearns' diagrams show the limestone transected by innumerable dikes merging to form a continuous intrusive body at depth. Assuming that the layer of limestone exists, then, provided the dikes made room for themselves by stoping, the limestone thus removed must have been assimilated in the magma. Judging from the nature of the dikes exposed in the shallow parts of Hawaiian rift zones, this is less probable than that the dikes were injected into fissures opened by lateral distension, caused either by external forces or by wedging action of the magma. But if so, it is entirely probable that a considerable amount of limestone remains as thin slices between the dikes, and thus is available for assimilation if and when the main magma body reaches that level. The upward advance of the main magma body may take place by stoping; or it may progress by a gradual heating of the overlying material and the injection of innumerable dikes, until the entire complex of dikes and intervening septa becomes hot, plastic, and eventually part of the magma body. Although either process of rise may be dominant, both probably operate together to some extent. No matter which is dominant, undoubtedly the magma body gradually becomes enlarged upward. Stearns appears to believe that the magma body reaches the level of the supposed limestone layer quite early in the magmatic cycle. However, if surficial eruption commences at or soon after the time when the abyssal magmatic wedge starts upward through the suboceanic crust, the magma chamber should not reach the limestone horizon until rather late in the principal volcanic period. Particularly if the magma body advances by heating and mobilization of the roof rock, as Stearns believes, such heating should continuously increase throughout the period of active volcanism and frequent dike injection, and the magma chamber should reach its maximum upward extension at or near the end of that period. The limestone horizon, at the former ocean floor, should be fairly close to the upper limit of extension of the main magma body; thus, the main magma body should encounter and assimilate the limestone near the end of the main period of volcanism. The time of appearance of the picrite-basalts of the ankaramite type, near the end of the period characterized by frequent eruptions of dominantly primitive basaltic lavas but at or near the beginning of the period of eruption of the andesites, conforms well with this picture. Inoculation of the magma with lime may even, as suggested by Daly, have stimulated differentiation and the production of the andesitic types.

The assimilation of limestone occurs, under this hypothesis, much earlier than the eruption of the nephelinic rocks. However, the effects of limestone assimilation might well continue over a long period of crystallization leading finally to the production of the nepheline basalts and associated rocks. Although limestone assimilation does not appear essential for the production of the nephelinic rocks, it is entirely possible that it may have taken place; and it would furnish a ready explanation for the distinctly higher content of lime in these lavas as compared with the parent olivine basalt.

Conclusions

All other Hawaiian rocks are believed to have been derived from olivine basalt magma largely through crystal differentiation. Basalts were formed by the removal of olivine crystals and picrite-basalt of the oceanite type by the addition of olivine crystals to olivine basalt magma. The andesites and trachytes are residual salic differentiates resulting from the removal from olivine basalt magma of crystals largely of calcic plagioclase, hypersthene, diopside, and olivine. Selective resolution of basaltic wall rocks may also have played a part, though probably small. Picrite-basalt of the ankaramite type owes its origin largely to enrichment of olivine basalt magma in crystals of olivine and calcic plagioclase and a little magnetite, together with the addition of a little lime by the

assimilation of limestone. The nepheline basalts, melilite-nepheline basalts, nepheline basanites, and picrite-basalts of mimosite type are believed to be the products of nearly complete crystallization of a cafemic submagma derived by the assimilation in olivine basalt magma of sunken crystals of olivine, pyroxene, and calcic plagioclase, and possibly a little limestone.

It is not intended to imply that the origin of these rocks in the manners indicated is conclusively proved, but only that they could have originated in these ways. However, it is believed that, in view of present knowledge of Hawaiian lavas, such origins are the most logical of any thus far suggested.

During the early stages of Hawaiian volcanism olivine basalt, basalt, and picrite-basalt of the oceanite type must exist together in the magma reservoir, probably arranged gravitationally. At a later stage the interbedding of olivine basalts, andesine andesites, and picrite-basalts of the ankaramite type from the same volcano indicate the simultaneous coexistence of these types as eruptible magmas. All three are extruded in fairly large volume, suggesting that all three come from the main magma chamber. Some such gravitative arrangement may occur in the magma chamber as illustrated in Figure 10, dikes tapping different levels in the chamber and bringing to the surface lavas of different composition. Liquid immiscibility is not implied, but merely a difference in specific gravity of the liquids, and sufficient stagnation of the magma body so that the layering is not destroyed by stirring. The boundaries are, of course, not sharp. There is a complete gradation from one magma layer to the next, reflected in a similar compositional gradation among the erupted lavas.

The occurrence of trachytes at Hualalai Volcano and in the Waianae Range of Oahu in small bodies interbedded with olivine basalts, without accompanying andesites, suggests their development in small parasitic magma chambers. On West Maui and Kohala volcanoes, on the other hand, they are associated with and grade into much more voluminous oligoclase andesites. In general, the trachytes are the latest lavas erupted at these volcanoes, except for very late post-erosional eruptions on West Maui, and they are probably the extreme salic differentiates of the main magma chambers.

The intermingling of nepheline basalt, melilite-nepheline basalt, nepheline basanite, "linosaite," picrite-basalt of the mimosite type, and olivine basalt, during the post-erosional eruptive period on Kauai and eastern Oahu, probably indicates the coexistence of many small residual magma reservoirs in various stages of crystallization, under varying conditions of temperature and pressure.

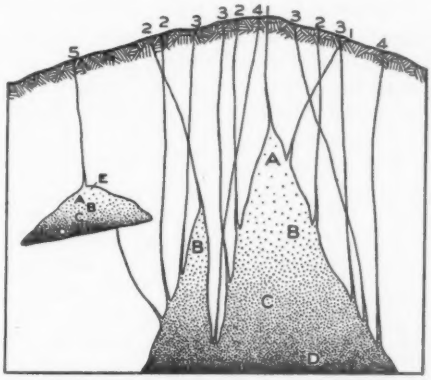

FIGURE 10.—DIAGRAM ILLUSTRATING A POSSIBLE MANNER OF NEARLY SIMULTANEOUS ERUPTION OF LAVAS OF DIFFERING COMPOSITION FROM THE SAME MAGMA CHAMBER

A, andesite; B, basalt; C, olivine basalt; D, picrite-basalt; E, trachyte.

Olivine basalt recurs throughout the volcanic history, from the oldest stages to the youngest. It is absent among the late (though not post-erosional) lavas of West Maui and Kohala, and among the posterosional lavas of the Koolau Volcano on Oahu, where, however, the "linosaites" closely approach olivine basalt in composition. Some of the latest (posterosional) olivine basalts of Niihau and Kauai are richer in titania than those of the primitive shield volcanoes. In general, however, from oldest to youngest, the olivine basalts are indistinguishable from each other in composition. This constancy of composition suggests the continuous or repeated addition, throughout the history of Hawaiian volcanism, of fresh olivine basalt from the same underlying source, either through fissures reaching the substratum directly, or by

means of fresh injections into the main magma body underlying each volcano.

Comparison with other Central Pacific Islands

The rock types of the Hawaiian Islands are characteristic, with minor variations, of the rocks of nearly all the volcanic islands of the central Pacific Ocean. Present knowledge of the distribution of types of volcanic rocks in central Pacific islands is summarized in Table 14. In the table certain minor rock types are included with their near relatives or synonyms. Thus "tahitites" in the Society Islands, tephrites in the Society and Tubuai islands, and "luscladites" in the Tubuai Islands are included with the basanites; and "ankaratrites" in the Society, Cook, and Hawaiian islands are included with the nepheline basalts. Intrusive rocks are included with their extrusive heteromorphs. Monchiquite and "shoshonite" are listed, despite their limited occurrence and small volume, because of their distinctive compositions which will not readily permit their inclusion with any of the other rock types listed. The "shoshonites," which are known only in Moorea, in the windward group of the Society Islands, are of special interest because, with the exception of a nepheline trachyte also from Moorea (Iddings and Morely, 1918, p. 114, no. 8), they are the only rocks of the central Pacific volcanoes in which potash exceeds soda.

Table 14 indicates that the only islands in the central Pacific in which the rocks are markedly different from all the rest are Easter Island and possibly the near-by Sala-y-Gomez. The presence there of dacite and rhyolite, in addition to olivine basalt, andesine andesite, and trachyte, suggests the affinity of the Easter Island Volcano with the calc-alkaline (andesitic) volcanoes of the circum-Pacific belt. Rhyolite has been recorded from the Marquesas Islands by Barth (1931b, p. 525), but this rock appears to resemble the quartz trachytes of Tutuila, Samoa, rather than the true rhyolites. The high submarine ridge on which Easter Island is situated may well consist of sialic rock. This is supported by the observed speed of earthquake waves reflected from points on the Easter Island ridge northeast of Easter Island (Fig. 11, C), reported by Gutenberg and Richter (1941, p. 37, Fig. 4), who regard the Easter Island ridge as probably marking the southeastern edge of the major Pacific Basin. It is so shown in Figure 11, in which the hypothetical boundary of the Pacific Basin has been swung eastward to include the rock assemblages of mid-Pacific type in the Galapagos Islands. Knowledge of Cocos Island is insufficient to justify extending the boundary to include it in the mid-Pacific province, especially in view of the report by Gutenberg and Richter of continental structure north of the Galapagos Islands. Gutenberg and Richter suggested that such areas of mid-Pacific rock types as those at the Juan Fernandez and San Ambrosio islands represent small areas of Pacific crustal type within the boundary of the continental area. The distribution of earthquakes and of reflection points indicating continental rocks (Gutenberg and Richter, 1941, Fig. 4) appears, however, to suggest that the Easter Island ridge may be a long, probably thin tongue of continental rock projecting southwestward from the Peruvian bulge of South America (Fig. 11). The calc-alkaline affinities of the Easter Island rocks probably result from the incorporation into the magma of some proportion of sial.

Lacroix (1927) has divided the islands of the central Pacific into three groups on the basis of lithology. He differentiates a nephelinic series in which practically all of the rocks contain modal or normative nepheline; an intermediate or mixed series containing some feldspathoidal rocks but with nonfeldspathoidal rocks dominant; and a series without nepheline, even in an occult state. The only island which definitely falls into the nephelinic series, on the basis of present knowledge, is Tahiti; and the present writer is inclined to suspect that even Tahiti, when sufficient knowledge is gained, will fall into the mixed series. The rest of the Society Islands, and indeed most of the mid-Pacific islands (including the Hawaiian group), appear to belong to the mixed series. Easter Island and Sala-y-Gomez belong to the series without nepheline. Lacroix places Mangareva and Pitcairn islands, in the eastern part of the Tuamotu arc, in the nepheline-free series, but it appears more probable to the writer that with more complete knowledge they also will be found to belong in the mixed series.

Burri (1926, p. 172–178) has divided the

TABLE 14.—TYPES OF VOLCANIC ROCKS IN CENTRAL PACIFIC ISLANDS

Name of archipelago or island	Olivine basalt	Basalt	Picrite-basalt, oceanite type	Picrite-basalt, ankaramite type	Picrite-basalt, mimosite type	Limburgitic basalt	Basanite	Monchiquite	Nephelite basalt	Melilite-nephelite basalt	Andesine andesite	Oligoclase andesite	"Shoshonite"	Trachyte	Phonolite	Dacite	Rhyolite	Source of information
Hawaiian	X	X	X	X	X		X		X	X	X	X		X				This paper
Clipperton														X				Teall (1898)
Caroline (eastern)*	X	X						X	X	X				X				Yossii (1937); Bridge (1948)
Cocos	X														?			Chubb (1933)
Galapagos	X										X	X		X				Washington and Keyes (1927); Richardson (1933)
Marquesas		X	X	X		X					X			X	X			Lacroix (1927); Chubb (1929, 1930)
Wallis	X											X						Macdonald (1945b)
Samoan	X	X	X	X		X					X	X		X				Macdonald (1944d)
Tuamotu†	X		X	X							X							Marshall (1918); Lacroix (1927)
Society (Leeward group)	X		X	X								X		X				Lacroix (1927); Stark and Howland (1941)
Society (Windward group)	X	X	X	X		X	X	X			X	X	X	X	X			Lacroix (1927); Williams (1933)
Cook	X	X		X					X		X				X			Lacroix (1927); Marshall (1909, 1927, 1930)
Tubuai (Austral)		X	X	X			X				X				X			Lacroix (1927); Smith (1927)
San Felix and San Ambrosio		X					X							X				Willis and Washington (1924)
Easter (Rapa Nui)	X										X			X		X	X	Lacroix (1927); Bandy (1937)
Juan Fernandez	X	X		X			X							X				Quensel (Daly, 1916, p. 336)

* The western Caroline, or Palau, Islands contain rocks of continental types, in contrast to the eastern Carolines which belong to the central Pacific province.
† Including Pitcairn Island.

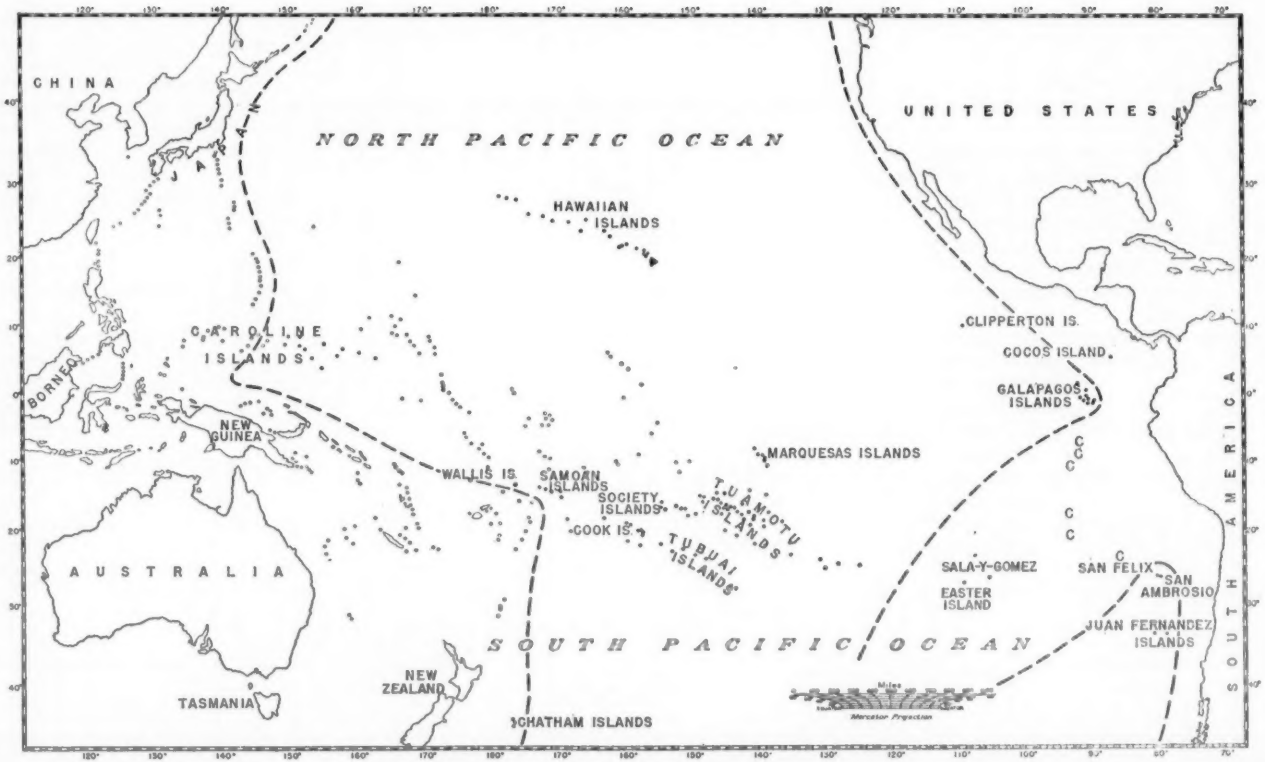

FIGURE 11.—MAP OF THE PACIFIC OCEAN

Showing position of the islands listed in Table 1 (solid dots), the hypothetical boundary of the Pacific Basin proper (dashed line), and reflection points of earthquakes indicating continental structure (*C*). The open circles represent coralline islands or islands composed of continental-type rocks, not listed in Table 14.

rock series of the mid-Pacific province into two types: the Hawaiian, and the Tahitian. On the basis of Niggli numbers, the Tahitian type is characterized by higher *alk* and *al*, and lower *fm* and *c*, than the Hawaiian. As examples of the Hawaiian type, Burri lists the Hawaiian Islands, the Leeward Society, Moorea in the Windward Society Islands, the Samoan Islands, Juan Fernandez, San Felix, and San Ambrosio. The only central Pacific island listed as belonging to the Tahitian type is Tahiti itself.

Except for Tahiti, Easter Island, and possibly Sala-y-Gomez, the rocks of all the mid-Pacific volcanoes are probably essentially similar. On the basis of existing analyses and rock descriptions, the rocks of Tahiti are distinctly more alkalic than those of the others. This is shown also by the alkali-lime index, which for Tahiti is about 46 (Williams, 1933, p. 41), as compared with about 50.5 for the lavas of the Samoan Islands (Macdonald, 1944e, p. 1356) and 54 for the Hawaiian Islands. Until Tahiti is completely and carefully mapped geologically, it will not be certain that the existing analyses adequately represent the entire volcanic structure. However, present indications are that the apparent more strongly alkalic character of the Tahitian rocks, and possibly those of the Cook and Tubuai islands, is real. Phonolites occur, and even the trachytes of the Society Islands generally contain a little occult or even actual nepheline; whereas the trachytes of the Hawaiian, Samoan, Galapagos, and Marquesas islands contain a small amount of quartz, occult in most, but appearing in the mode of some Samoan examples.

Other minor differences occur. Certain Tahitian rocks are richer in titania than most rocks of the other islands. However, titania-rich rocks are present among the post-erosional lavas of Kauai and Niihau, in the Hawaiian Islands, and elsewhere. Possibly the titania-rich lavas of Tahiti also belong to a group more recent than the major volcanic structure. Probably only in the Hawaiian and parts of the Samoan islands are the stratigraphic and structural relationships of the various types of lavas really well understood. Even in Tahiti, where the various rock types, especially those of unusual composition, have had very thorough petrographic study, much more field work will be necessary for a complete understanding of the spacial and time relationships (Williams, 1933, p. 34). Until such relationships are known it is risky to draw fine distinctions on the basis of minor compositional variations, many of which can be duplicated within the boundaries of single archipelagoes or even single islands.

Notwithstanding minor differences, the striking feature of the rocks of the central Pacific islands is their similarity over the entire province. General conditions must have been essentially uniform throughout the area. The parent magma must have been practically the same in composition, and is probably represented by the olivine basalt which constitutes an overwhelming share of the major edifices of all the well-exposed and well-known mid-Pacific volcanoes.

Williams (1933, p. 46) has suggested that in Tahiti assumption of an alkali basalt (basanitoid) as the parent magma offers the least difficulties in the derivation of the other rock types, but believes any more definite conclusions to be unwarranted. It appears unlikely, however, that a parent magma uniform over a large part of the Pacific basin would be replaced only at Tahiti by a parent magma of different type. It is more likely that the same general parent magma has been affected at Tahiti by some different process, or more probably to a different degree of effectiveness by one of the same processes which have operated in the other volcanoes. Conditions and processes of assimilation and differentiation must also have been very similar throughout the mid-Pacific province. The somewhat greater abundance of normative and modal quartz in Samoan trachytes, as compared with those of Hawaii, may have resulted from such minor differences as shallower magma chambers, with resultant lower pressures and separation of a greater proportion of olivine at the expense of hypersthene (Macdonald, 1944e, p. 1360). The greater abundance of alkalic rocks, and their greater deficiency in silica, in the Society and neighboring islands, as compared to the rest of the mid-Pacific, may have resulted from the assimilation of a greater proportion of limestone there. However, it should again be emphasized that these differences are minor. The

Showing position of the islands listed in Table 1 (solid dots), the hypothetical boundary of the Pacific basin proper (dashed line), and selected points of earthquakes indicating continental structure (○). The open circles represent coralline islands or islands composed of continental-type rocks, not listed in Table 14.

characteristic features are similarity of rock types throughout the area, and by implication essential uniformity of petrogenic processes.

The olivine basalt-trachyte associations typical of the central Pacific province are characteristic likewise of volcanic islands rising from the depths of other oceanic basins. They are not, however, wholly restricted to the oceanic areas, but occur also in some continental areas, such as southeastern Australia and the Otago Peninsula of New Zealand (Edwards, 1935). In contrast, except at Easter Island, calc-alkaline andesite-dacite-rhyolite series such as are typical of the continental border zones are unknown in the mid-Pacific. Edwards points out that the olivine basalt-trachyte associations in continental areas were erupted in districts which had experienced no major orogeny for very long periods previous to the eruptions, whereas the calc-alkaline eruptions typically accompany and follow orogeny. The calc-alkaline magmas probably are the result of assimilation of sialic rocks by olivine basalt, with accompanying differentiation; whereas the alkaline olivine basalt-trachyte magmas are the result of differentiation alone, without appreciable assimilation of sialic rocks. Orogeny may promote assimilation of sial by preheating the sialic rocks, as suggested by Edwards (1935, p. 14), but assimilation of sial is probably at least partly the result of down-bowing of the sialic crust, during orogeny, into zones of high temperature. In the central Pacific, where the sialic shell is absent or extremely thin, orogenic downwarping has brought no sial into zones of elevated temperature where large-scale assimilation can occur, hence the calc-alkaline andesite-dacite suites are absent. Similarly, under the Atlantic and Indian oceans the sial is so thin that orogeny has not in general depressed it into the highly heated levels of great assimilation. In the eastern border regions of the Pacific, the areas of mid-Pacific rock types such as Juan Fernandez and San Ambrosio, which apparently lie within the boundaries of the South American sialic continental platform, may be areas of thin sial or merely areas in which orogeny had been absent for long periods before the eruptions.

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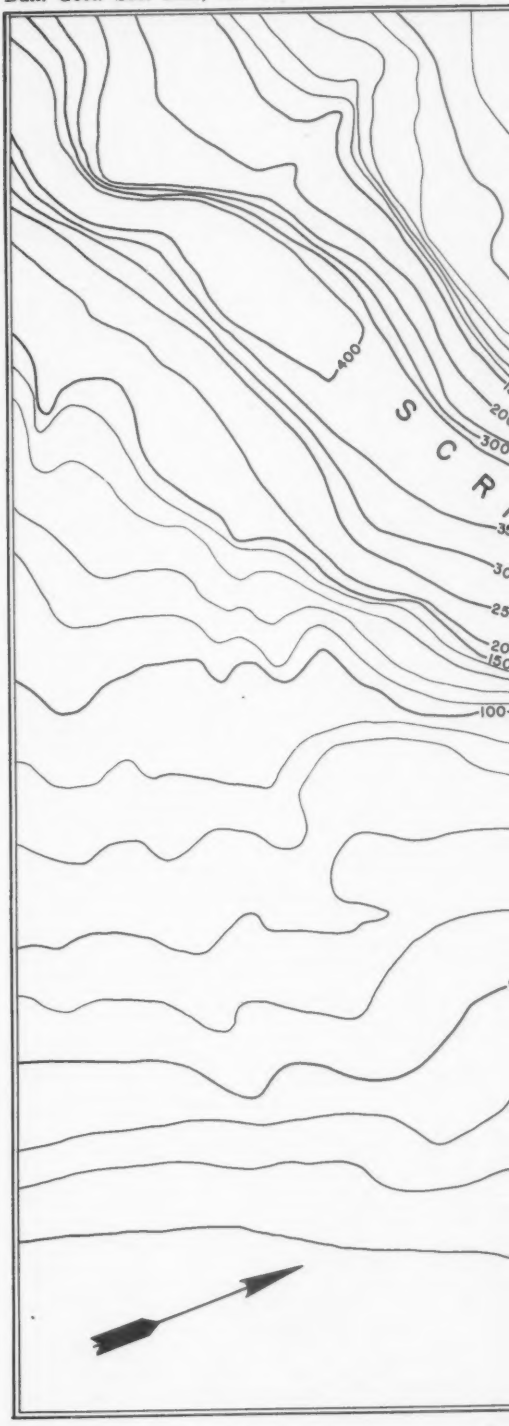

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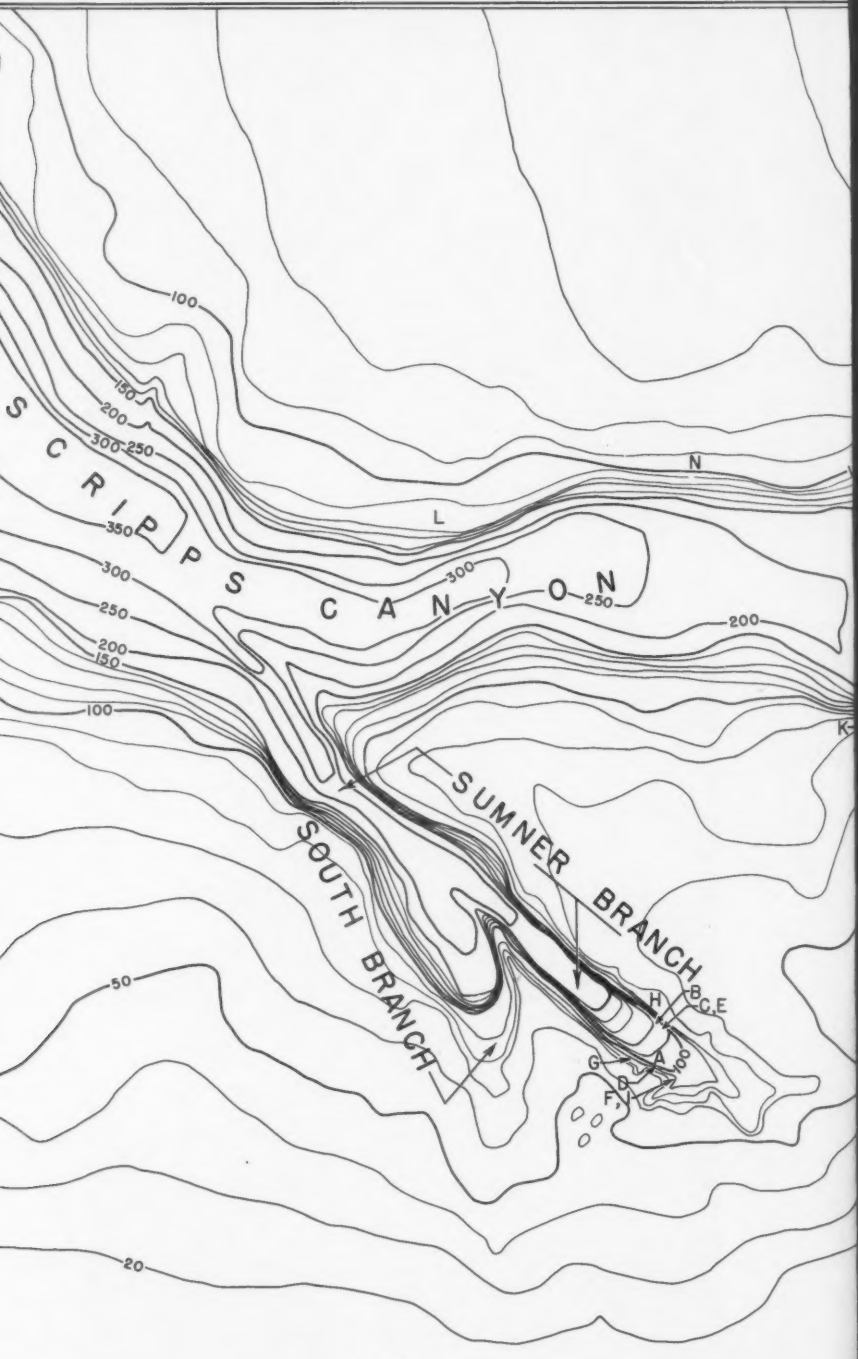

CONTOUR MAP OF HEAD OF SCRIPPS C

though map is based on thousands of soundings it probably does not show
ures of the sea floor in this area. Note vertical cliffs and the flat but stee
ers give location of photographs. Note also change in contour interval at

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TERRESTRIAL TOPOGRAPHY OF SUBMARINE CANYONS REVEALED BY DIVING[1]

BY FRANCIS P. SHEPARD

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ABSTRACT

To obtain more detailed information concerning the nature of submarine canyons, diving operations were carried on in the heads of two typical canyons in the La Jolla area. Frank Haymaker, who dove under the writer's telephonic direction, described and photographed features which show that these canyons closely resemble the adjacent land canyons. Haymaker found vertical and even overhanging rock walls, narrow tributaries entering either at grade or as hanging valleys, and sediment-covered canyon floors. Cliffs of alluvium with layers of cobbles were discovered at the head of one of the canyons. In none of the 62 dives, extending over a period of a year, was there any evidence of significant density currents nor of the effect of any other strong currents moving down the canyon floors. The evidence suggests that the canyons were excavated by streams and that the heads are being filled but

[1] This work represents in part the results of research carried out by the Office of Naval Research, the Hydrographic Office, and the Bureau of Ships under contract with the University of California.

that the fill is removed from time to time by mud flows. Although the dives did not go deeper than 190 feet in the canyons, evidence is presented to show that the topography could not be due to moderate lowering of sea level.

Introduction

In October 1946 Frank Haymaker, a naturalist and former Navy diver, made the first dive into the submarine canyons of the La Jolla area (Fig. 1), 12 miles north of San Diego. Landing on the east wall of Scripps Canyon just above a vertical cliff, he described through the two-way telephone system a scene of extraordinary

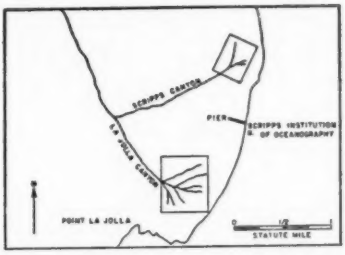


Figure 1.—Axes of the La Jolla Canyons
Areas of Plate 1 (head of Scripps Canyon) and of Plate 8 (head of La Jolla Canyon) indicated.

beauty as well as of great interest to an investigator of submarine canyons. The remarkable clarity of the water, giving a visibility of about 30 feet on the bottom, gave great promise for photography in the canyon heads. However, this water clarity proved to be most exceptional; despite 55 intervening dives, no such water was seen again until January, 1948, when a 9-day period offered ideal conditions. Prior to this time dives yielded some photographs and indicated the points in the canyons where photographs would prove particularly worth while if the water cleared. Also this earlier period gave an opportunity to get the photogrraphic equipment working smoothly and to learn the most efficient methods. Shortly after the January period of clear water ended, the operation was forced to an end by a northwest blow which hit with little warning and caused the loss of the 38-foot boat used in the dives and the greater part of the diving equipment.

The photographic portion of the diving operations was only one of several investigations. For example, the interpretation of the photographic results and of the telephonic descriptions given by the diver required a large amount of soundings to form a general picture of the areas in which the diver was operating. Also a study was made of rocks, marine sediments, and alluvium collected by the diver from canyons. Other activities such as current measurement and coring of the canyon floors were carried on only to an extent sufficient to warrant passing mention in the present report.

The investigation of the canyon heads by diving operations is not without economic application. In recent years, jetties for three California harbors have been built directly into the heads of submarine canyons. Relative to beach erosion and to the fate of the harbor jetties, it has become important to determine both the exact nature of the canyon heads and the processes operating in the canyons.

Acknowledgments

The diving investigation of the La Jolla canyons was made possible by funds from the Penrose bequest of The Geological Society of America and from Navy contracts with Scripps Institution. The skill and zeal of Frank Haymaker in getting the photographs, many of them in most difficult diving terrains, and the detailed descriptions by Haymaker of the areas were of the greatest possible help. Haymaker had majored in zoology at the University of California at Los Angeles and during his association with Scripps Institution of Oceanography made a considerable study of geology in order to improve the accuracy of his description of geological features which he encountered on the bottom. The photographic enlargements of the 25-mm films used in the submarine photography were largely the work of Haymaker, and his skill in "dodging-in" the important features added greatly to their value. The dive tending by D. B. Sayner and R. L. Wisner soon became so professional that the operations moved with far greater smoothness than could normally be expected from amateur dive tending. Acknowledgment should also be made of the effectiveness of the Ewing submarine camera devices which were used for most of the work. These were equipped with handles and a rack for flashlight bulbs, the latter devised by Haymaker. Thanks are due Dr. Roger Revelle for many helpful sug-gestions in the manuscript.

Nature of Operations

Diving into a narrow submarine canyon which has vertical and even overhanging walls is a hazardous undertaking. It requires careful investigation of depths and surface currents prior to anchoring, and the laying of both bow and stern anchors in such places that they will not slide down into the canyon. The location for the dive is predetermined by a combination of ranges and sextant angles, a buoy with short scope is dropped and then the anchors are lowered and the boat is jockeyed into position by manipulation of the anchor lines at either end of the boat. Care had to be taken to avoid anchoring at points where rip currents were in evidence since the changing currents in the outer ends of the rips cause considerable shift in position of the boat, endangering the diver's lines.

The writer directed the dive operations through two-way telephones. When Haymaker reached the bottom he would report on the state of visibility. When the visibility was fair but too poor for photography, the diver would walk around the bottom looking for features which might permit future photographs. Such features were located as closely as possible by the amount of the life line which was out and the direction in which the diver's bubbles were rising. When interesting rocks were discovered, a skiff was rowed over to the diver and a sledge hammer and a chisel lowered so that he could break off the rocks and send them up in a bag. Breaking off rock under the surface is not easy since it is difficult to swing a hammer, but in general the diver was able to obtain fragments. Haymaker obtained samples of the fill on the canyon floor by driving in pipes. The personnel at the surface would pull out the pipe, and Haymaker would tie a cover over the pipe end to hold in the sample. It was difficult to obtain long cores because of the presence of hard packed sand. A more effective coring means is being built for future operations.

Directions during the dives were obtained from a compass worn on Haymaker's wrist. At frequent intervals Haymaker gave the direction and relative velocity of the bottom current and the nature of the bottom surge. When the current was too weak to be determined from the movement of floating debris, it was observed by allowing a line, with the end weighted so as to have zero buoyancy, to float out from the diver. The use of special instruments for further current work on the canyon floors is being planned.

In bottom photography an attempt was made to use a camera mounted on a tripod, but the setting up of the tripod frequently stirred up so much sediment that photography was very ineffective. Also the time required to set up the tripod greatly interfered with the making of a complete photographic record. By holding the Ewing camera case in his hands Haymaker could keep ahead of his own murk and could take 10 times as many photographs as were possible with a tripod. The location of each photograph was noted as closely as possible, and the diver's description was recorded.

Murkiness in Canyons

The diving operations into the canyon heads were greatly hindered by the murkiness of the water. Day after day the water had just enough silt or plankton so that the tributaries, the overhanging rocks, and the vertical cliffs described by the diver would fail to register on the films. An extreme contrast was noted on a day when the water on the shelf at the side of the canyon was so clear that we could see bottom in 55 feet of water. At the canyon rim the diver could see only about 10 feet. Looking down into the canyon Haymaker found that the murk was general.

The turbidity in the canyon heads is attributable in part to suspended sediments, derived from the washing away of the shale and mud at the base of the wave-cut coastal cliffs, and in part to plankton. The murk in Scripps Canyon was almost always bad after spring tides when the waves reached the base of the coastal cliffs shoreward from the canyon. It was, of course, also bad after all periods of large waves. The effect of the runoff after the very mild rain storms which occurred during the 15-month period of diving operations was not very great. So far as could be told, most of the mud moved out over the surface and soon disappeared without settling into the canyons. The mud washed from the cliffs and carried out by rips gravitated slowly toward the canyon, and

the gentle back-and-forth currents appeared to be just sufficient to keep particles in suspension along the canyon bottom.

It had been supposed that in summer after the end of the stormy season the water in the canyons would clear, but unfortunately when the waves had gone down the plankton had so increased that the water was as murky as ever. At times the plankton were in layers, making it possible to operate in clear water below the cloudy layer.

Finally, in January 1948 the plankton had disappeared, and the waves were small so that the water at last cleared. The period of clear water in all of the canyon heads began at a time when the bottom currents had apparently ceased.

Scripps Canyon

General Topography

In 1934 Scripps Canyon was surveyed in considerable detail by wire soundings. At that time echo sounding was completely out of the question because (1) the sonic sounders available could be used only on large vessels which could not possibly maneuver into the near-shore positions at the head of the canyon, (2) the machines lacked recording devices which are essential to pick up details, and (3) the walls proved to be so steep and the base so narrow that the wide cone of the sound "beam" would bring in echoes from walls alone rather than from the floor of the canyon. Today, the first two difficulties have been removed, but the third still remains although soon Bottom Scanners (Russell, 1946) may give us correct profiles along the axis of the canyon by sound. Meantime, we have been using a recording echo-sounding device which attaches to the 38-foot buoy boat used in the diving operations. With this instrument, considerable mapping has been possible for the canyon heads and less precipitous portions of the walls. The instrument shows the presence of tributaries, many of them only gullies, which are difficult to pick up by the much more laborious process of wire sounding. Furthermore, echo soundings make it easily possible to prove the actual existence of any feature by running repeat lines along ranges on the same day. This is important since occasional false echoes may come either from schools of fish or from some temporary defects in the electronic system.

The accompanying map (Pl. 1) of the head of Scripps Canyon and of its two main tributaries is by no means perfect, but it is probably at least as good as most surveys of land canyons. Recent discoveries indicate that there are more tributaries and more vertical walls than had been supposed. Probably further soundings will show that there are even more of these features. Examination of the contours indicates that the canyon has three main tributaries at its head. The southern two branches extend shoreward to depths of about 16 feet and to within about 500 feet of the average low tide line on the beach. They are both located almost directly off Sumner Canyon (Pl. 2), a profound valley which cuts into the 300-foot sea cliffs in the vicinity. The inner portions of these tributaries are gently sloping, sand-covered valleys which have floors and walls with so little relief that it is not detectable by a diver. They both show a sudden change in gradient at axis depths of about 75 feet for South Branch and 50 feet in Sumner Branch. Beyond this the gorgelike character of the canyon begins, and rock appears on the walls. At depths of 100 feet, Haymaker determined that the floor itself was comparatively flat and covered with sediment despite its high gradient of about 25°. The two tributaries join at a depth of about 230 feet; they are separated higher up by a knife-edge ridge. Wire soundings were most difficult to obtain near the juncture of these two canyons because of the precipitous slopes and the large error resulting from the lead line being even a few degrees out of plumb.

The northern head of Scripps Canyon extends in to within 800 feet of the cliffed shore line, terminating near the mouth of Callan Canyon, the most impressive land canyon of the district (Pl. 2). The head of this main branch of Scripps Canyon has a depth of about 50 feet. Like the other branches it deepens precipitously, dropping at least 50 feet in 100 feet beyond the 57-foot depth. So far as is known, all of the outer canyon has an incised inner gorge with clifflike walls which extend to the bottom of the canyon. In addition to Sumner Branch, which comes in at grade, the southeast wall of the main canyon has a series of tribu-

AIRPLANE VIEW OF SEA CLIFFS AND LAND CANYONS INSIDE SCRIPPS CANYON

Rip currents which supply fine suspension sediments to the waters within the canyon base are shown to the right.

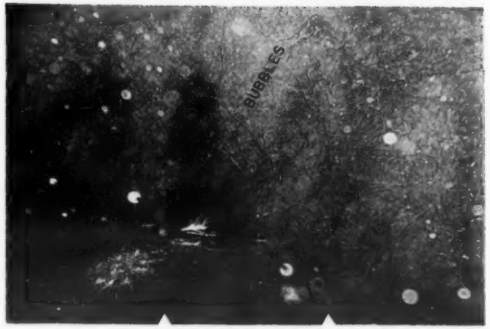


FIGURE 1. COLUMNS OF GAS BUBBLES RISING FROM FLOOR OF SUMNER BRANCH
The bubbles are the milky zones rather than the round objects which are due to reflected light from suspensions in the water. Depth 100 feet. Location at A, Pl. 1.

FIGURE 2. SHALLOW GULLY AT BASE OF CLIFF ON NORTH SIDE OF FILLING OF SUMNER BRANCH
Depth 100 feet. Location at B, Pl. 1.

FIGURE 3. CONCRETIONARY LENS PROJECTING FROM VERTICAL WALL ON NORTH SIDE OF SUMNER BRANCH
Depth 100 feet. Location at E, Pl. 1.

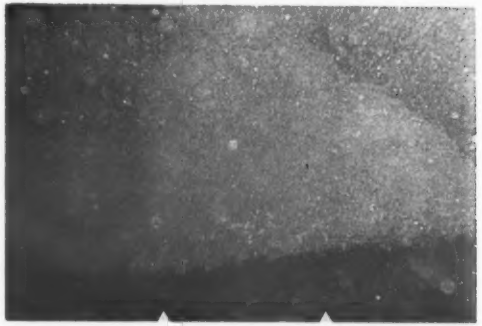


FIGURE 4. LARGE OVERHANGING MASS OF ROCK ON SOUTH WALL OF SUMNER BRANCH
Photo taken from floor of tributary valley. Projecting mass is directly above portion of cliff shown in Pl. 5, fig. 1. Depth 80 feet. Location at F, Pl. 1.

tary valleys many of them hanging. The northwest wall, however, is largely free from tributaries in the inner portion.

Dives into Sumner Branch

Thirteen dives were made into Sumner Branch. The early ones indicated the existence of remarkable features, but great difficulty was found in obtaining good photographs. The dives on January 21 and 23, 1948, into the head of Sumner Branch were sources of great excitement to both diver and surface party since even the area at the canyon bottom, which always had been blanketed in murk, was remarkably clear, and the features which had been so hard to describe previously now became vividly apparent. Because of faulty film the pictures on January 21 did not come out, but those taken on January 23 are satisfactory. Haymaker followed the same route as previously. His description over the phones during these two dives gave a very clear idea of the area, agreeing with the photographs in almost every respect. The diver landed on the canyon floor which he described as sloping seaward at about 20° and being comparatively smooth. The floor was thickly covered with eel grass and kelp. Bubbles[2] were seen rising in large quantities (Pl. 3, fig. 1). We could, of course, see them at the surface. The floor proved to be about 40 feet across. On the north side, Haymaker observed a gully which looked like a gutter on the side of a road. It was about 2 or 3 feet across and 8–10 inches deep (Pl. 3, fig. 2). The gully floor lacked all signs of current ripples. It could be traced for at least 30 feet along the base of the cliffs on the north side. Nothing like it was seen on the other side or in the center. The canyon floor on these occasions was fairly firm so that Haymaker sank into it only a few inches. He had described the same floor as very soft during a previous dive on August 28, 1947. The explanation can be assumed to be either compaction during the interim or the sliding out of the top of the fill.

The contact between the fill on the floor and the canyon walls in this area is shown in Plate 4. The rock walls rise vertically from the floor. Haymaker described the rock stratification as horizontal in this particular section, which allows one to judge the angle of contact with the fill. The nine available photographs show that this angle ranges from a slight back slope to 45° down canyon with the majority sloping about 15° to 20° down canyon. Since the floor was described as fairly uniform in its down-canyon slope, either the floor material in slumping down canyon has left irregular contacts along the side or material falling from above has built small fans at the base. The existence of such fans was not supported by most of Haymaker's observations although he thought he saw one example below a hanging valley.

Probably the canyon fill comes largely from the canyon head. If the fill came from sliding of sediment over the vertical walls, fans should be abundant at the base of the cliffs.

The walls are well stratified (Pl. 4). Concretionary lenses project as overhanging masses (Pl. 3, fig. 3). Most of the large overhanging masses consist of resistant formations (Pl. 3, fig. 4). Prominent joints are indicated in some of the photographs. Pholad borings are shown in most of the pictures (Pl. 4) and were found in the rocks which were recovered. Unfortunately not many rock samples are available from this area, because emphasis was put on obtaining photographs during the January dives, and it was difficult to break off some of the projecting rocks on a previous occasion. The rocks from the top of the cliff are impure limestone with a high percentage of silt and sand grains in the insoluble residue. This rock is a member of the Eocene Rose Canyon shale formation (Hanna, 1926).

In the vicinity of the photographs, the cliffs are about 40 feet high. The vertical portion of the cliffs increases in height outwardly until it is over 100 feet near the juncture of Sumner and South branches. The slope above the cliffs is variable but in general not very steep. On the edge, there are projecting masses of rock partly covered with Gorgonia corals and other growths. Elsewhere small nicks or gullies cut the rock (Pl. 5, fig. 1). Above the rocky edge is a sand slope, which was always observed to be deeply ripple marked. At one place well out on the north rim the ripples terminate at the very wall of the canyon (Pl. 5, fig. 2).

At the head of the rock gorge three gullies come down into the canyon at steep gradi-

[2] The bubbles probably consist of methane gas. They were first observed in 1937.

ents on either side and enter at grade. A photograph was taken from the 30-degree sloping floor of one of these tributaries looking toward the main canyon (Pl. 5, fig. 3). The vertical character of the canyon wall is clearly indicated. Rock from the tributary wall proved to be shale despite the vertical slopes. Haymaker followed up this tributary to where it ended on the divide between Sumner and South branches. From there he walked west until he came to another narrower tributary which was about 20 feet across at the top. He followed this to a point where it terminated as a hanging valley on the edge of the clifff. Like the main valley, these tributaries were floored with eel grass and kelp mixed with fine sediment, mostly sand. Other tributaries cut the top of the cliff on the north side. No opportunity was found to trace these tributaries downward, so it is not known whether they enter the main canyon at grade or as hanging valleys. One of them had a gradient estimated at 45°.

On the ridge between Sumner and South branches, Haymaker described a sand floor with an undulating surface suggestive of sand dunes or giant ripples with height of 6–10 feet. Unfortunately, no photograph shows these features although several kodachromes showed small sand ripples in the foregound with only a hazy indefinite background.

Dives onto Walls of Scripps Canyon

Haymaker made 12 dives onto the walls of Scripps Canyon, all of them located near the break in slope which descends to the canyon floor. In the areas visited, the floor was of the order of 200 feet deep, and on most occasions the murk appeared considerably worse below the break in slope than above, so that it was not thought advisable to take the large risk of such a deep dive. Above the break the visibility was ordinarily better than at corresponding positions in Sumner Branch nearer the shore.

On the southeast side of the canyon, the sandy slope leading to the steep inner gorge is broken in places by rock outcrops. Near the edge the rock is quite continuous and is covered with a luxuriant growth of organisms. On the upper slopes where the depths varied from 75 to 120 feet, ripple marks were reported only in the bottom of a gully. These ripples crossed the axis of the gully and indicated a moderately strong current moving either up or down the axis.

The break in slope of the southeast side of Scripps Canyon is irregular. Gullies cut the upper slope at each locality visited on this side of the canyon. The only successful photograph of a gully (Pl. 5, fig. 4) shows a narrow ravine with irregular rock slopes largely covered with a growth of Gorgonia corals and other organisms. The gullies steepen at the edge to as much as 60°, and at least one of them terminates as a hanging valley. Along the edge of the cliff between the gullies, a considerable number of jutting rocks were observed. Haymaker photographed them by leaning out over the precipice and being supported by his life line.

Because of the extremely clear water at the time of the first dive in 1946, Haymaker was able to look down the cliffs for about 40 feet. He could see the layers of rock dipping down canyon to the southwest at an angle of about 10°. Measuring the cliffs with an inclinometer showed that they averaged about 80° but had distinct overhangs in many places. Farther out along the axis of the canyon the heights of these cliffs range up to about 250 feet. In the area where the dives were made the heights average about 100 feet. Unfortunately nothing is yet known of the contact between the base of the cliffs and the canyon floor. The only rock samples obtained from the southeast wall were sandstone considerably bored into by pholads, but soft shale was found above the break in slope.

The northwest wall of Scripps Canyon was the site of seven dives. These covered a length of 900 feet above the break in slope, and one extended across the shallow head of the canyon. Visibility was never very good during these dives so that only a few satisfactory photographs are available. The slope steepens toward the cliffs and becomes progressively more rocky. To the southwest, the slope above the cliff is covered with huge rocks which made it difficult for the diver to proceed. The growth of Gorgonia is most abundant in this area and is almost forestlike in appearance (Pl. 6, fig. 1). Growth extends out to the brink of the canyon. Farther up canyon to the northeast, the slopes are less rugged and have sand-covered terraces

FIGURE 1. CONTACT OF NORTH WALL OF SUMNER BRANCH WITH FILL ON CANYON FLOOR
Rock face is vertical. Note pholad borings. Depth 95 feet. Location at C, Pl. 1.

FIGURE 2. CONTACT OF SOUTH WALL OF SUMNER BRANCH WITH FILL ON FLOOR
Note that angle is quite different from that of Figure 1. Great abundance of kelp and eel grass also illustrated. The fish is a Blacksmith (Chromis punctipinnis). Depth 100 feet. Location at D, Pl. 1.

FLOOR AND VERTICAL WALLS OF SUMNER BRANCH

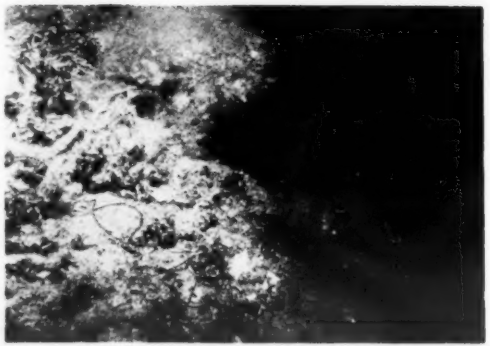


FIGURE 1. SMALL GULLY IN UPPER EDGE OF CLIFF ON SOUTH SIDE OF SUMNER CANYON
Depth 60 feet. Location at G, Pl. 1.

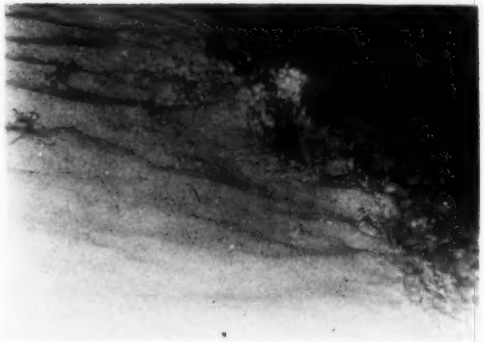


FIGURE 2. RIPPLE MARKS IN SAND EXTENDING TO EDGE OF STEEP SLOPE OF NORTH SIDE OF SUMNER BRANCH
Depth 70 feet. Location at H, Pl. 1.

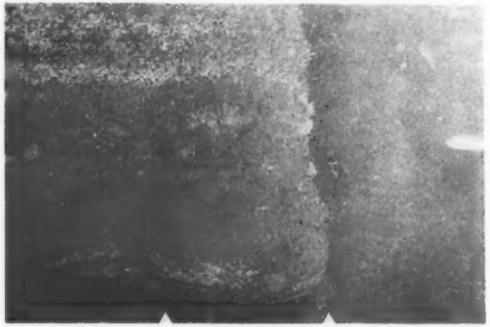


FIGURE 3. VERTICAL CLIFF
Picture taken from steeply sloping floor of tributary near its juncture with main valley of Sumner Branch. The fish is a Blacksmith (Chromis punctipinnis). Depth 75 feet. Location at I, Pl. 1.

FIGURE 4. GULLY LEADING INTO MAIN HEAD OF SCRIPPS CANYON ON EAST WALL
Note extensive Gorgonia growing on the rocks. Depth 90 feet. Location at J, Pl. 1.

FIGURE 1. ROCK TERRACE ON WEST WALL OF SCRIPPS CANYON
Note large size of the Gorgonia in the background. The fish (Sebastodes serriceps) and the holothurian (Stichopus californicus) were close to the camera. Depth 115 feet. Location at L, Pl. 1.

FIGURE 2. ROCK SLOPE AND SAND TERRACE ON WEST WALL OF SCRIPPS CANYON
Terraces provided easy progress along the steep canyon wall. Depth 90 feet. Location at M, Pl. 1.

TERRACES ON SIDE OF SCRIPPS CANYON

FIGURE 1. RIPPLE-MARKED SAND AT HEAD OF LA JOLLA CANYON
Note sand dollars (Dendraster excentricus) which are turned on edge and live in great abundance in this locality and at the head of Scripps Canyon. Depth 38 feet. Location at A, Pl. 8.

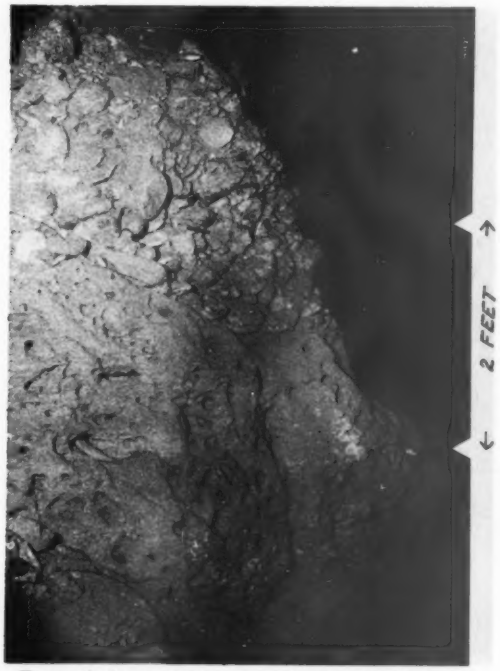


FIGURE 2. ALLUVIAL MATERIALS EXPOSED ON RIDGE BETWEEN TWO GULLIES AT HEAD OF LA JOLLA CANYON
Note abundant holes in sandy alluvium below cobbles. Gobies (Coryphopterus nicholsii) can be seen in the holes, and a sea anemone on the shoulder at the top of the picture. Depth 95 feet. Location at B, Pl. 8.

RIPPLED SAND AND ALLUVIAL FORMATIONS AT HEAD OF LA JOLLA CANYON

(Pl. 6, fig. 2), along which it was easy for Haymaker to progress. Walking northeast during a dive near the canyon head, Haymaker came onto a sand slope and crossed the canyon head above the rock gorge. Here the slopes are gentle. No ripples were reported from this sand area where the depth was 80 feet.

The break in the slope below the terraces is as spectacular on the northwest side of Scripps Canyon as in the other localities described above. Despite considerable walking along the edge, no good example of a gully was observed. Projecting rocks and slight rock benches were found. Also at a number of places the cliffs appear to be distinctly overhanding.

La Jolla Canyon

General Topography

Although the head of La Jolla Canyon is practically devoid of rocks, the topography is strikingly irregular (Pl. 8). At its head the canyon breaks off precipitately. In many places there is a descent of 100 feet in an equal horizontal distance. Vertical cliffs of small magnitude are common. There are large tributaries on either side, particularly to the north where two arms extend for over 1000 feet diagonal to the shore. One southern tributary is known as "Sea Lion Gulch" because during 6 months of the year a large colony of sea lions has chosen the head of this gully as their rendezvous. Ridges between tributaries are either knife-edged or rounded. The tributaries widen as they approach the main valley. The floor below the point where the tributaries join to form the main valley is comparatively flat although steeply sloping. The valley floor is about 200 feet wide at depths of around 180 feet below sea level but narrow at greater depth.

Dives in the Gullies at the Canyon Head

Twelve dives were made into the central portion of the canyon head, mostly among the gullies which abound in this area. Gentle-walled valleys are found in the shallow water with rippled sand slopes supporting an abundance of echinoids (Pl. 7, fig. 1). These gentle valleys terminate abruptly in steep-walled gullies. Haymaker described the descent of one of these gullies as being like sliding down an open drain. Vertical walls are common on the sides of these gullies (Pl. 7, fig. 2). Ridges between the gullies are narrow and irregular (Pl. 9, fig. 1). At the base of some of the cliffs there are distinct talus cones (Pl. 9, fig. 2).

The cliffs and steep walls of the gullies consist of alluvium, similar to the alluvium of the Pleistocene (or Recent) sea cliffs south of Scripps Institution. For the most part this material is a moderately compact, poorly sorted loam with a striking absence of marine organisms. It has numerous holes in which small fish (known as gobies) are living (Pl. 7, fig. 2). Some of the cliffs toward the center of the canyon head have beds of cobbles (Pl. 7, fig. 2; Pl. 9, fig. 2), which are predominantly of crystalline rock as are the alluvial cobbles on land in this area. Lack of stratification and jumbled appearance indicate that in both cases the cobble beds are the result of flood deposits. The marine cobble beds have a few boulders up to about 1 foot in diameter.

The bottom of the gullies is covered with muddy sand containing considerable kelp and eel grass. The proportion of plant debris increases outwardly to depths of about 130 feet beyond which it decreases. Some of these algae are also found on the shelves above the cliffs, and in places stalks of kelp hang over the cliffs looking like strands of rope.

At an axis depth of 185 feet, the canyon floor was different in appearance. Eel grass and kelp were not at all common (one ball of eel grass was found), and the floor was sand with only a trace of the silt which was so abundant in shoaler water. So far as could be seen in the restricted field of a photograph which was limited by the usual murk, the surface of the sand was bumpy, although the roughness was not sufficient to impede progress over the bottom. The consistency of the bottom material was described by Haymaker as being "lumpy but firm".

Sea Lion Gulch and the Hanging Valley

Herds of sea lions apparently like to have a meeting ground on the ocean at some point which they can identify even in the fog. Dr. Carl L. Hubbs and J. L. McHugh of Scripps Institution have observed that the sea lions

of the herd which gathers in winter off La Jolla show a decided tendency to congregate in the same geographic locality 1000 feet out from the shore. We made detailed soundings at this locality and found that it was at the approximate head of one of the southern tributaries of La Jolla Canyon at a depth of only about 30 feet. On one day the water was so clear that using a waterglass we were able to observe the sea lions swimming around the canyon head. That their location is approximately the same during a fog was determined by listening to the barking of the sea lions during the fog from shore points directly inside the gathering ground. Therefore, they must use the submarine topography to reach their rendezvous.

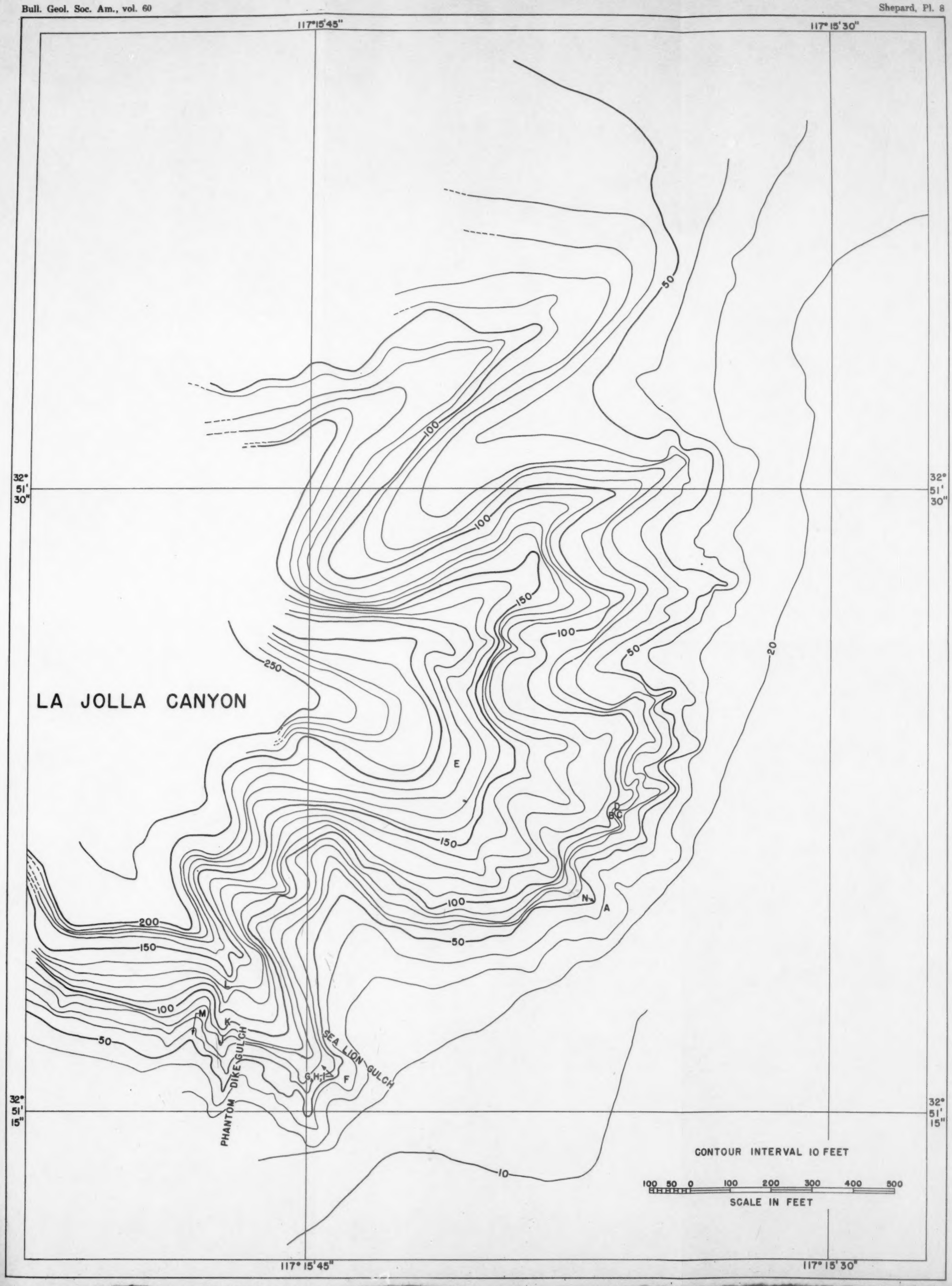
Dives into Sea Lion Gulch showed that it has comparatively gentle walls but is definitely a gully with a narrow axis which could be followed by the diver. The axis of the valley has an abundance of shells and sand together with cobbles up to about 10 inches in diameter. Pens from squid are very common, being undoubtedly the remains of the food of the sea lions. Some of the cobbles may have been sea-lion gastroliths. The fact that the pebbles and cobbles include a surprising abundance of soft calcareous shale fragments, 30 per cent by count, the remainder being of crystalline rock, indicates that the material has not been subject to great transportation. Rocks taken by Emery (1941) from the stomach of a dead sea lion contained four rounded shale fragments out of a total of 23 pebbles. On the other hand, pebbles on the local beaches have almost no shale. The collections from Sea Lion Gulch contain rocks weighing up to 5 pounds, probably too large for gastroliths.

Following down Sea Lion Gulch to a depth of about 65 feet, Haymaker came to a cliff with a vertical drop of 5 feet (Pl. 10). The floor widened here so much as to give the impression that the gully did not extend farther down; later this was found not to be the case. The cliff which cuts the Gulch extends beyond the confines of the inner gully and can be traced in either direction for a total length of about 50 feet. It could be seen to die out into the even slope of the canyon side. Apparently there are two recesses into the cliff marking the position of two separate gullies which are cut off by the cliff. Cobbles were seen in Sea Lion Gulch both at the top of the cliff and on the floor below (Pl. 10 fig. 2). The cliff contains the same poorly sorted alluvium as that of the main head of La Jolla Canyon. Holes in the alluvium house abundant fish larvae, and an octopus was seen in one hole.

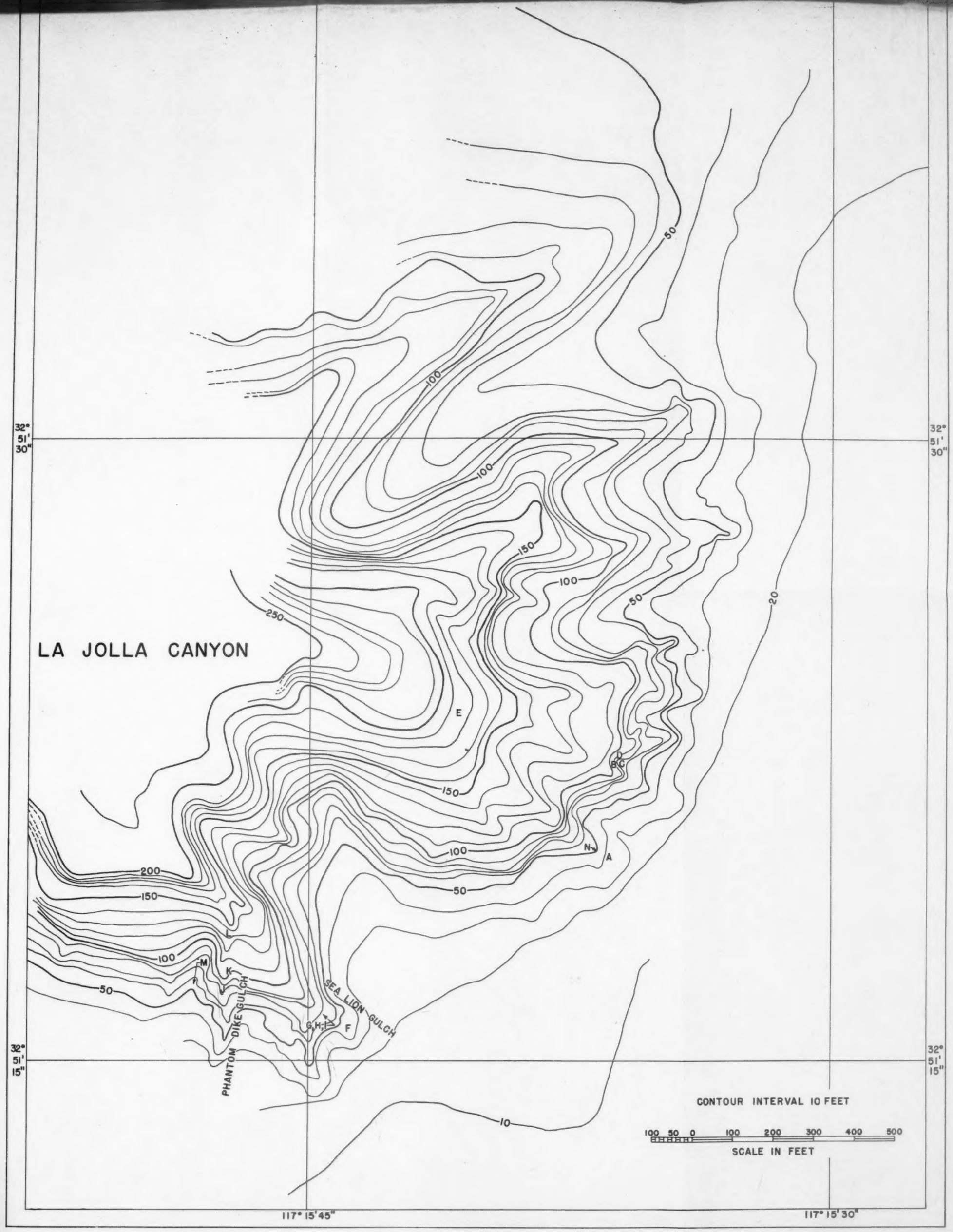
Phantom Dike Gulch

A small steep-floored tributary enters Sea Lion Gulch from the south near the point where the latter enters the main canyon. This tributary descends 100 feet in approximately 220 feet giving it an average gradient of 24°. In an early dive into this gulch Haymaker encountered the only rock outcrop of all the diving operations in the head of La Jolla Canyon.[3] He described the rock as forming a vertical wall about 1 foot high which extended up the northeast side of the gulch with all the appearance of a dike. He broke off a large piece of the rock, which surprisingly, proved to be a well-cemented, highly calcareous, poorly sorted sandstone with clear lines of stratification parallel to the flat sides, which according to Haymaker, formed the sides of the vertical wall. The camera was out of order during this dive, but the locality was carefully marked and was given high priority for future work. Returning during the remarkably clear water of January, 1948, Haymaker made two descents along the axis of the gully looking at the sides with the greatest of care, but despite the good visibility he was unable to find a trace of the rock. Careful study of the fixes and ranges leaves no doubt that Haymaker was in the same gully, but a search for the rock was made in the neighboring gullies, to no avail. The bringing up of the rock broken from the wall leaves no doubt of its having been there; the rock had disappeared, apparently having been covered by slides of the sediment from above despite its projection of almost a foot above the surroundings. The occurrence of a vertical formation in this area of horizontal or slightly inclined rocks is puzzling. An origin as a sandstone dike is somewhat unlikely because of the stratification. Possibly a layer from the walls had toppled over and sunk into the sediment in this position.

[3] Rock has been dredged farther out in the canyon (Emery and Shepard, 1945).



CONTOUR MAP OF HEAD OF LA JOLLA



CONTOUR MAP OF HEAD OF LA JOLLA CANYON

Despite abundance of soundings contours had to be generalized in places and the map fails to show the numerous small cliffs found by diving operations. Letters give location of photographs.

Phantom Dike Gulch, named for the elusive rock wall, was found to be floored by sand and mud with an abundance of kelp and eel grass in the upper portions but the gulch floor has sand and mud alone lower down. Going down the axis, the walls steepen and are vertical in places (Pl. 11, fig. 1). At the lower end of the valley, in about 140 feet of water, the slope flattens, and the gully enters the main valley at grade. At this place a large block of alluvium about 4 feet across and 3 feet high (Pl. 11, fig. 2) was found resting where it had apparently stopped in a slide down the steep valley floor.

An examination of the relatively flat-topped ridges between gullies in this area showed an interesting assortment of organisms (Pl. 12, fig. 1). Along with echinoids and holothurians there was a large amount of kelp. Cobbles and angular rocks lie scattered over the surface.

South Wall of La Jolla Canyon

To the west of Phantom Dike Gulch, the tributaries become less common. Two dives were made onto the shelf at the edge of the canyon. The terrace above the break is covered with coarse sand having an abundance of broad-leaf eel grass (Zostera). There is an abrupt change in slope at the brink, and a comparatively smooth mud and sand slope extends down at an angle of approximately 45°. The sediment proved to be largely very fine sand with abundant mica and a moderate amount of silt. The abundance and variety of marine life were in contrast to the paucity in canyon floor samples. In a few places there were small pockets which appeared to be the result of slumping. In walking along this slope Haymaker started small slides in the muddy sand. Much of the disturbed material went into suspension. The muddy water thus formed started down the very steep slopes but was ·oon carried clear of the bottom, rather than ontinuing as a bottom-flowing current which ·me geologists would have expected. The gen- al smoothness of the wall on these steep slopes certainly surprising. However, no great thick- ss of sediment may exist here because the mery-Dietz coring device penetrated only a v feet of sediment near the base of the slope ore encountering a mass hard enough to smash the nose. It seems likely that these steep sediment slopes are only temporarily in repose and are subject to occasional slides.

Bottom Current Observations

Very few quantitative data relative to bottom currents were obtained from the dives because of the necessity of concentrating on the photography and general description of the canyons. Haymaker did, however, note the amount of surge and the net transport of floating objects. When possible he estimated the velocity. The currents observed were found to reverse in direction at least once during the course of each dive (during a period of half an hour to an hour). That is, if the current was flowing down canyon at the beginning of the dive, it was later observed flowing up canyon. Of the total number of current observations, 47 per cent showed currents moving down canyon, 30 per cent up canyon, 7 per cent across the canyon, and 16 per cent cent showed no current. The slight excess of down-canyon currents may be due to suspension (density) currents set up by the diver's progress over the bottom. The currents measured in submarine canyons by instruments have included approximately as many up-canyon movements as down (Shepard, 1948a, p. 62, 63). On one occasion Haymaker estimated that the current velocity was almost half a nautical mile an hour, but the rest of the time the velocity was much lower. There was no evidence that the currents were stirring up the bottom, and the water was remarkably clear on the day when the current was estimated as being as high as half a knot (Pl. 11). If this velocity existed, it must have been much slower at the very bottom. The feebleness of the currents probably accounts for the general absence of ripple marks on the canyon floors. Except in very shallow water ripples were only found in one place, a hanging tributary to Scripps Canyon. The feeble currents also explain the silt in the sediment samples on the canyon floors in contrast to the clean sand which was usual on the terraces of the canyon walls. In the latter localities the shoaler depth allows more wave stirring and winnowing of the fine sediments. However, relatively clean sand was discovered at the deepest dive in 185 feet of water, in a

locality which was below the areas of muddy and kelp-covered bottom.

A sediment trap (Pl. 12, fig. 2), which has fine mesh socks placed so that their open ends face in four directions, was placed by Haymaker in the head of La Jolla Canyon. Two days later it was brought in, and the sediment content of the socks was weighed. It was found that the bags which received sediment carried by up-canyon currents had 64 grams, while those exposed to down-canyon currents had 38 grams.

During the first part of the period of excellent water clarity, in January 1948, the currents were almost missing. However, during the latter part of this period the currents were observed moving down canyon with unusually high velocities and without the usual reversals. During this same period Dr. Walter Munk of Scripps Institution reported that the long-period waves (averaging about 15 minutes) which had been appearing regularly on the long-period wave (tsunami) recorder at Scripps Pier had disappeared. It will be of interest to look for further correlations between the canyon conditions and these long-period waves.

Comparison of Submarine and Land Topography

It might be supposed that soundings of the submarine canyons would yield much better data than diving operations for comparing the canyons of the sea floor with those on land. While it is true that soundings are the only method of bringing out the major canyon features, it is almost impossible to pick up small but important details by them. For example, we knew that there were very steep slopes in Scripps Canyon from the wire soundings, but it took an on-the-spot observer to find that these slopes included actual vertical and even overhanging cliffs, hanging valleys, and steep tributaries entering at grade. The diving operations were confined to the upper portion of the La Jolla canyons, but there seems to be no evidence from the soundings that the canyon characteristics change appreciably until much greater depths are reached. Therefore, it seems reasonable to assume that the dives have yielded us evidence of the detailed character of a considerable portion of typical submarine canyons.

The sea cliffs north of Scripps Institution have a series of canyons which are cut below the surrounding terrace to depths comparable to those of the submarine canyons where diving operations took place. Also the land canyons are cut into the same types of Eocene rocks as the branches of Scripps Canyon. Furthermore, the heads of the land canyons are cut into alluvium which resembles that of the head of La Jolla Canyon and its tributaries. The comparisons between the two types of canyons are accordingly of considerable significance.

The general resemblance of the drainage pattern of Scripps and La Jolla submarine canyons to the adjacent land canyons has been discussed elsewhere (Shepard and Emery, 1941, p. 59). The canyon trends show the same right-angled bends and even the same two principal axial directions. The comparison extends even to the details.

A general view of the land canyons is provided by an airplane photograph (Pl. 2), which may be compared with the generalized map (Fig. 1), the contour maps (Pls. 1, 8), and with the block diagram of Sumner Branch which Donald Sayner has prepared from data from the diving operations and the soundings (Fig. 2). The general resemblance between the land and the sea-floor topography seems to be unmistakable. The many branched heads found in both of the submarine canyons are also found in all of the land canyons. Both the land and the submarine canyons have many more tributaries in the upper reaches than lower down. Stretches of straight walls found in the land canyons (Pl. 2) compare with the ungullied northwest slope of Scripps Canyon.

Vertical cliffs like those of the submarine canyons are not uncommon in the land canyons (Pls. 13, 14). Tributaries enter either at grade or as hanging valleys in the land canyons and sea canyons alike (Pl. 13). Steps are found along the course of the land gullies (Pl. 13, fig. 2) just like the steps in Sea Lion Gulch. Jutting rock masses project from the cliff faces in the land canyons (Pl. 14, fig. 1) as do the jutting rocks of the submarine canyons. Projecting spurs are found in the alluvium along the course of the land valleys (Pl. 13, fig. 3) very much like the spur at the head of La Jolla Canyon (Pl. 9, fig. 1).

There are, on the other hand, some con-

FIGURE 1. FLATTENED PORTION OF RIDGE
Above Pl. 7, fig. 2. Compare Pl. 13, fig. 3. Larger fish, a California sheepshead (Pimelometapon pulchrum), and the smaller fish, a goby (Coryphopterus nicholsii) were abundant in this location. Depth 75 feet. Location at C, Pl. 8.

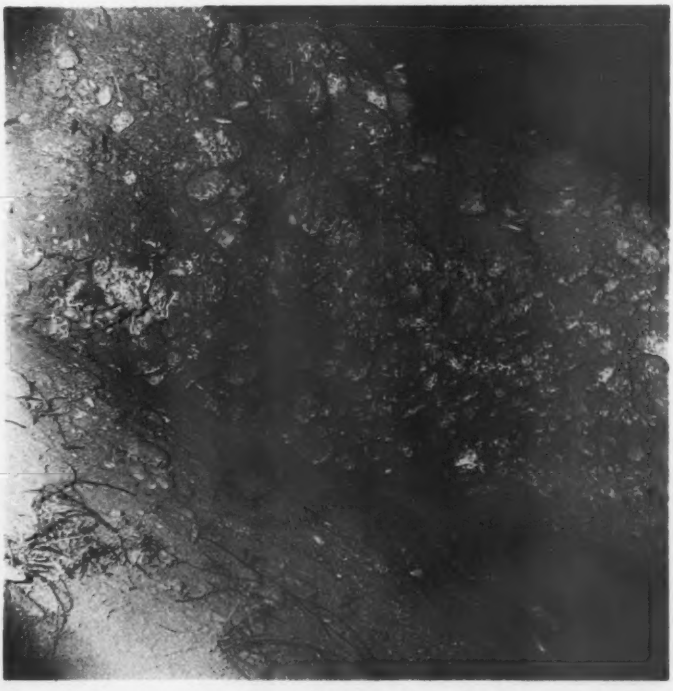


FIGURE 2. CONTACT OF COBBLE BED AT HEAD OF LA JOLLA CANYON WITH TALUS SLOPE
Note jumbled mass of cobbles indicative of flood deposition. Depth 90 feet. Location at D, Pl. 8.

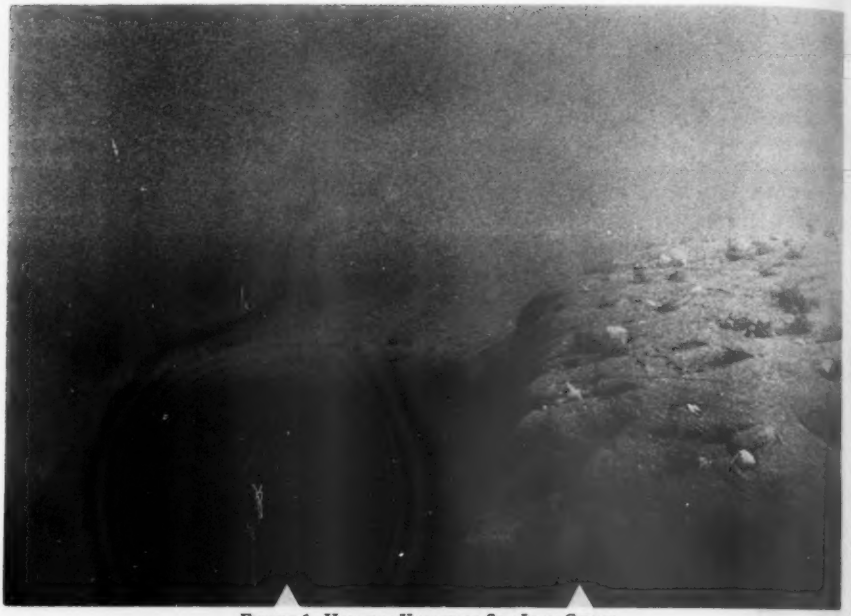


FIGURE 1. HANGING VALLEY IN SEA LION GULCH
Compare with Pl. 13, Fig. 2. Depth 65 feet. Location at G, Pl. 8.

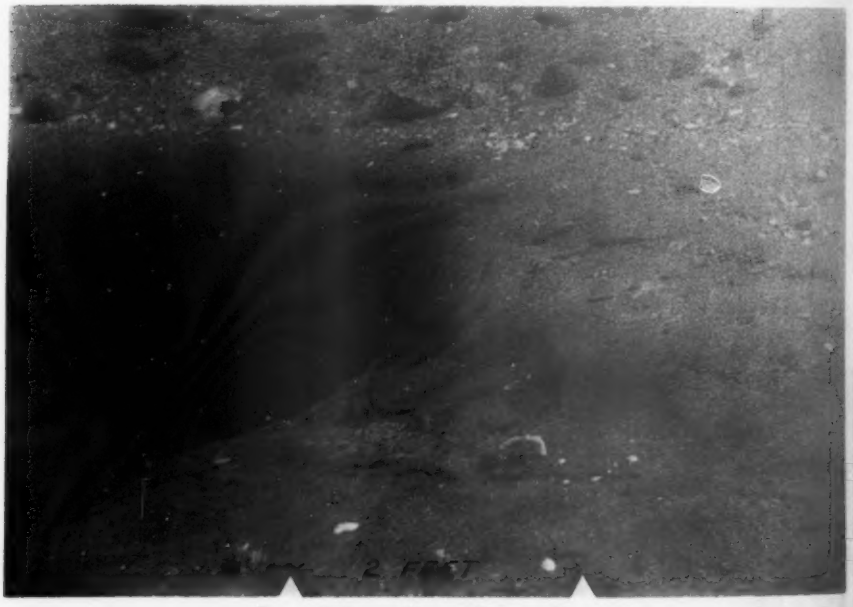


FIGURE 2. CLOSER VIEW OF HANGING VALLEY
Cobble-strewn floor of the valley above the 5-foot drop shown in background. Depth 65 feet. Location at H, Pl. 8.

HANGING VALLEY IN SEA LION GULCH

trasts between the land canyons north of La Jolla and those of the sea floor. The land canyons have extremely narrow notches at their in slope at the rim, and vertical walls extending down to the bottom. If this land gorge were submerged and sediment allowed to accumulate

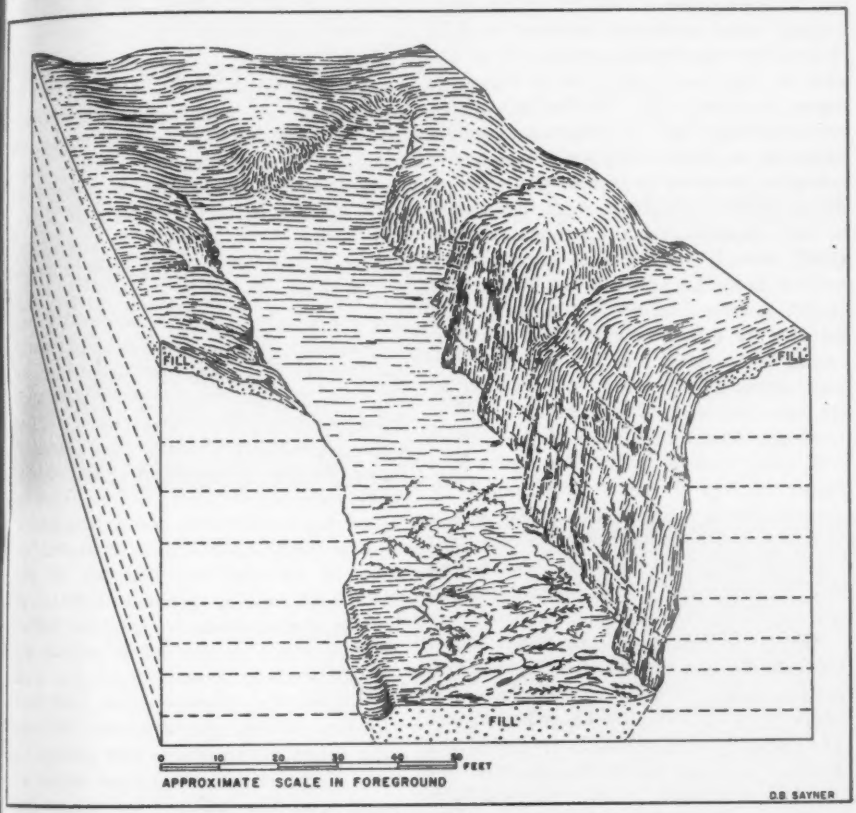


FIGURE 2.—BLOCK DIAGRAMS OF HEAD OF SUMNER BRANCH

The various features described for the dives in Sumner Canyon are illustrated. No vertical exaggeration intended. Diagram constructed by D. B. Sayner.

base just wide enough for a man to squeeze through. These have not yet been reported from the two principal La Jolla submarine canyons, although some of the tributary valleys are almost as narrow and have equally steep walls. The alluvial cliffs in the land canyons are very much fluted by erosion developing columnar structure. Such fluting has not been found on the ocean bottom.

Some of the tributaries of the Grand Canyon show a remarkable resemblance to the head of Scripps Canyon. Fig. 2 of Plate 14 shows such tributaries, an inner gorge with abrupt change in it, a flat floor similar to that in the submarine canyon might develop.

ORIGIN OF THE LA JOLLA CANYON HEADS

Suspension (Density) Currents

So much has been written showing how out of line with the facts are the hypotheses of canyon origin by faulting, by tsunamis, and by artesian spring sapping[4] that it seems unnecessary

[4] Points cited are as follows:
Against faulting, Shepard and Emery, 1941, p. 110–11.
Against tsunamis, Kuenen, 1947, p. 45; Shepard

to repeat these as yet unanswered discussions. Certainly none of the features discovered by the diving operations support these interesting but unsatisfying hypotheses. Density currents, more properly called suspension currents, are still credited by a considerable number of scientists with at least contributing to an important degree to canyon origin. The fact has always been overlooked that the suspension currents which are so clearly developed in reservoirs are quite powerless to prevent deposition of fine material at the points where they should be most capable of cutting gorges. Kuenen (1947) recently attempted to apply the hypothesis to the La Jolla canyons. Unlike Daly (1936), Kuenen considers that the suspension currents are still in operation and thus can extend the canyons into the present shore. Since almost all the canyons (except those off the east coast of the United States) extend in to shoaler depths than those of the −40 to −60 meter level which Daly assumed existed during the time of canyon excavation, Kuenen's modifications would seem to be essential to bolster this hypothesis. Kuenen considered that canyons could be cut by the density flows provided they lacked sharp turnings and waterfall steps along the course of the canyons.

The diving operations provide new direct evidence. Nothing reported by Haymaker indicated that suspension currents have had anything to do with the La Jolla canyons. The sharp turns and waterfall steps thought by Kuenen to be contrary to the hypothesis are definitely established by Haymaker's pictures and observations. Unfortunately for most of the diving operations, suspension currents failed to clear the canyons of the suspended mud which settled into them after being carried out from shore by rip currents. During the murkiest dives, when suspension currents should have been moving water down canyon, the currents showed an alternation of up- and down-canyon movement.[5] The fills on the canyon floors testify to the failure of suspension currents to keep the canyons clean at least under ordinary conditions. The cutting of vertical and overhanging cliffs consisting of rocks so compact that they were difficult to break is certainly not to be expected from suspension currents.

Mud Flow-Landslide Hypothesis

All recent writers have considered landsliding or mud flows along the ocean floor as an important process. Evidence of these mass movements is becoming well established from the studies of depth in the heads of the La Jolla canyons. The small trench in the fill of Sumner Branch (Pl. 3, fig. 2) might be considered evidence of a mud flow.

Could such movements produce deep elongate canyons with characteristics like those of the canyons off La Jolla? Given an appreciable slope at the edge of the continental shelf, what would happen if this inclined surface remained submerged over a long period of time? We can assume that much material would creep or slide down the slope and, if the underlying material were poorly consolidated, broad amphitheater-like indentations would develop. If the edge were underlain by well-consolidated rocks, as appears to be the case off La Jolla, changes should be so slow that millions of years would be required to produce any appreciable effects. Even if irregularities in the outer slope were developed, how could these turn into elongate narrow canyons with deep inner gorges and tributaries with patterns like those of the stream-cut canyons on the adjacent land? This difficulty has not yet been explained by those favoring the landslide hypothesis. Furthermore, the canyons apparently lack the basin depressions so common in landslide topography.

Excavation by Streams

The descriptions of the landward heads of the La Jolla submarine canyons and comparisons with adjacent land canyons seem to provide conclusive evidence that the process which formed the land canyons was also responsible for those on the sea floor. Such differences as exist are easily explained by sedimentation on the canyon floors and by the absence of rain wash on the canyon walls. After submergence, the extremely narrow gorges of the land canyons

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and Emery, 1941, p. 120–21; Shepard, 1948a, p. 59, 60, 240, 241.

Against artesian spring sapping, Rich, 1941; Daly, 1942; Kuenen, 1947, p. 43–44; Shepard and Emery, 1941, p. 115–116.

[5] In dives at the head of Monterey Canyon, where the water is very muddy, weak alternating currents were also reported by Haymaker.

FIGURE 1. VERTICAL WALL OF ALLUVIUM IN LOWER PART OF PHANTOM DIKE GULCH
Note fish (Coryphopterus nicholsii) protruding from holes in the alluvium. Depth 100 feet. Location at K, Pl. 8.

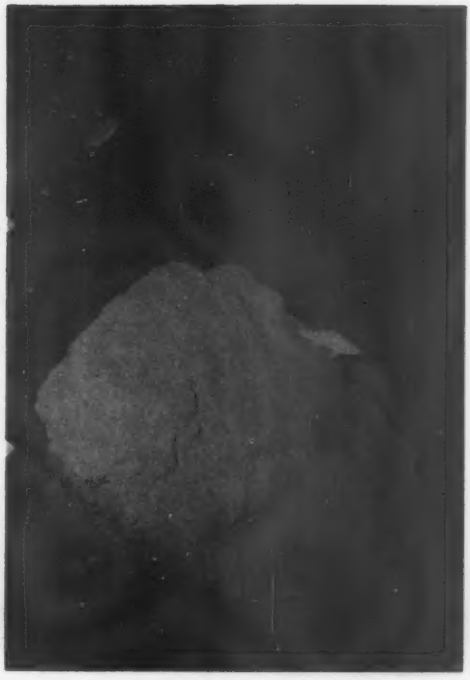

FIGURE 2. LARGE LUMP OF CLAY AT MOUTH OF PHANTOM DIKE GULCH
Found at point where gulch entered wide portion of main canyon. Fish are kelp bass (Paralabrax clathratus). Depth 130 feet. Location at L, Pl. 8.

VERTICAL ALLUVIAL CLIFF AND LARGE CLAY LUMP IN LOWER PHANTOM DIKE GULCH

FIGURE 1. RIDGE WEST OF PHANTOM DIKE GULCH
Note abundance of life, including large holothurian and sea anemone, and the cobbles mixed in with the sand and kelp stalks. Good lighting in background comes from sunlight. Depth 58 feet. Location at M, Pl. 8.

FIGURE 2. NEW TYPE OF SEDIMENT TRAP IN HEAD OF LA JOLLA CANYON
Socks caught more sand coming from the down-canyon side than from the up-canyon direction. Depth 50 feet. Location at N, Pl. 8.

RIDGE BETWEEN GULLIES NEAR PHANTOM DIKE GULCH AND SEDIMENT TRAP IN HEAD OF LA JOLLA CANYON

FIGURE 1. AIRPLANE VIEW OF GLIDER CANYON CUT INTO SEA CLIFFS NORTH OF HEAD OF SCRIPPS CANYON
Note hanging tributary.

FIGURE 2. HANGING VALLEY ON LAND CUT IN ALLUVIUM
It is comparable to the hanging valley of Sea Lion Gulch shown in Plate 10.

FIGURE 3. SHOULDER IN ALLUVIUM AT HEAD OF CANYON CUT IN CLIFFS NORTH OF SCRIPPS INSTITUTION
Compare a similar feature on the sea floor shown in Pl. 9, fig. 2.

FEATURES OF LAND CANYONS ADJACENT
TO SCRIPPS CANYON

FIGURE 1. VERTICAL NORTH WALL OF GLIDER CANYON
Note projecting ledges which are comparable to Pl. 3, fig. 4.

FIGURE 2. TRIBUTARY TO GRAND CANYON WHICH APPEARS TO BE SIMILAR TO HEAD OF SUMNER BRANCH
Compare Figure 2. Airplane photo by Robert S. Dietz.

LAND CANYON AT LA JOLLA AND TRIBUTARY OF GRAND CANYON

would quickly fill with sediment, and this fill would be held in place by the walls. The fluting and columns of the alluvial land cliffs would tend to disappear after submergence. As submarine weathering proceeded, these features, which are due to running water, would not appear unless density currents of the type predicted by Daly and Kuenen were flowing over the sides.

Clearly, the major characteristics of the La Jolla marine canyons are those of the adjacent land canyons. The evidence that land canyons are the result of stream erosion is so conclusive that few geologists deny it. Unless oceanographers and marine geologists have overlooked some important process, it seems only reasonable to assume that the canyons which Haymaker has photographed on the sea floor off La Jolla were excavated by streams when the sea was lower or the land higher.

Significance of the La Jolla Canyon Investigations

A prominent geologist, after hearing of the evidence from the dives relative to the canyon heads, asked why it was significant since it was well known that the sea stood several hundred feet lower during the glacial stages, so that the excavation described could have been due to a moderate lowering. The answer to this good question is that lowering of a few hundred feet would expose the continental shelves which have gentler slopes off most places, including La Jolla, than the adjacent land. As a result these gentle slopes would become covered by fans rather than truncated by narrow gorges. One can easily imagine the result of exposing an undissected shelf off the sea cliffs shown in Plate 2. Every flood would help the construction of fans on the emerged sea floor. Canyons like those off La Jolla would be cut only at places where the lowered sea level exposed steep slopes or steep valley heads formerly covered by the sea.

Furthermore, if cliffs without canyons had been exposed by a 40–60 meter lowering of sea level the new gorges would necessarily terminate at that sea level, just as the canyons in the marine cliffs north of Scripps Institution today terminate near the present sea level. A study of the charts of the world shows no such canyons along any coast.[6] Canyons like those at La Jolla give no indication of sudden change in character at 40–60 meters below sea level nor, for that matter, at much greater depths. The close approach of the canyon heads to the shore at La Jolla is nothing unusual and is found in scores of other places around the world. Nor does the La Jolla area show any indication of recent diastrophic activity. Broad wave-cut terraces are in accord with present sea level (Shepard, 1948a, Fig. 60). A continental shelf of approximately normal depth lies off the coast everywhere in the La Jolla area except where cut by submarine canyons.

Possible Sequence in Development of La Jolla Canyons

Primary Deep Excavation

Despite the objections raised by some geologists to the idea that the continental borders once stood much higher than they do at present (or that the sea was formerly much lower) submarine geological operations and almost all surveys of the sea floor continue to supply evidence which is extremely difficult to interpret except by invoking major changes of level (summarized in Shepard, 1948a, p. 241–248, 332–335; 1948b). In addition to the river-canyon topography of the sea floor, the finding of rounded cobbles, flat-topped sea mounts, terraces, and shallow-water shells in water hundreds of fathoms deep strongly indicates submergence. The borings and geophysical prospecting of coral reefs have all yielded evidence which is convincing to most geologists that the coral-reef areas have undergone great submergence (Kuenen, 1947; Ladd *et al.*, 1948). We need not discuss here whether the world-wide changes indicated by these interlocking lines of evidence mean that the sea stood lower at one time or the continental margins, oceanic islands, and sea mounts stood higher, nor is it necessary to consider whether the changes took place at one time or at several times. The important point seems to be that, unless all this evidence is void, coasts have formerly stood relatively much higher than

[6] Shallow sea valleys with gentle gradients and gentle wall slopes like that off the Hudson should not be confused with submarine canyons.

they do now. The evidence for southern California is particularly complete (Shepard and Emery, 1941). Therefore, it is by no means unreasonable to assume that at some time, possibly in the very late Tertiary[7] or in the Pleistocene, not only the shelf but the steep upper slopes of the San Diego Trough off La Jolla were above water. At such a time canyons would have been cut into these exposed slopes. This canyon cutting must have been of relatively short duration, unless the narrow gorges of Scripps Canyon and of parts of La Jolla Canyon represent a development more recent than that of the main canyons.

Drowning of Canyons

If we assume further that the sea or the land returned more or less to its present level, the sediments washed out from the lands would have tended to fill the canyon heads. However, in places where the canyon gradients were relatively steep or where the sediment was sufficiently mobile, the canyons would have been maintained by occasional mud flows or landslides. Elsewhere, the rapidly shifting sands would have soon obliterated the canyon heads. The outer portions of the canyons may have been filled by the very mud flows which were maintaining the inner portions, although it is equally possible that the mud in the flows went into suspension and was ultimately deposited as a thin blanket in San Diego Trough.

Accompanying the filling of canyon heads, there would have been straightening of the coasts by a combination of estuary filling and sea-cliff cutting into the projecting headlands.

Shifting Sea Levels of the Glacial Stages

If the major cutting of the canyons had preceded some or all of the glacial stages, the swinging sea levels of the Pleistocene would have had a very different effect on the dissected continental shelf than on a smooth shelf. The lowered sea levels would have laid bare the steeply sloping inner portions of the unfilled canyons, and as a result erosion would have taken place at the canyons heads. At the same time alluvium was forming on the inner shelf but eventually the headward erosion of the canyon would have cut through this a[ll]uvium. The forming of the alluvial-wall[ed] gullies at the head of La Jolla Canyon [co]uld thus be explained as the product of this glacial stage erosion. Because the last rise in sea le[vel] at the end of the Wisconsin took place [on]ly 10,000 to 25,000 years ago, it is not surpris[ing] that these alluvial walls should be still [well] preserved. The vertical walls at the head [of] Scripps Canyon may also be quite recent or [at] least recently re-excavated.

It is evident that the headward erosion during the glacial stages of lowered sea level did not cut the present land canyons below the present sea level (Pl. 2). The absence of estuaries or filled estuaries may prove puzzling to some readers, particularly in view of the existence of the high gradients at the canyon heads at a distance of about 900 feet from the cliff. In view of the present slow rate of rock-cliff retreat in the area (Shepard and Grant, 1947), probably not more than a foot per century, the present cliffs have probably not retreated very far in the few thousand years since the last rise in sea level. Assuming the existence of the sea cliffs and the land canyons during the last glacial lowering, it can be further assumed that deposition would take place on the exposed shelf, particularly at the canyon mouths. The fans formed as a result of this deposition would cause the spreading into distributaries of the occasional flood waters emerging from the present land canyons and would thus have prevented the concentration of water into the unfilled heads of the exposed submarine canyons. Furthermore, the headward erosion in the exposed canyon heads would have been somewhat counteracted by the fan deposition at the mouths of the present land canyons. As a result, retreat may not have been very great during the time the canyon heads were exposed.

Maintenance of the Canyon Heads at the Present Day

Studies of sand movements along the shore show that there is enough shifting of sediment to allow the filling of the canyon heads, as they now exist, in a few decades. Also, shoaling found at the canyon heads during certain periods demonstrates that the canyons are receiving this sediment. Since the canyons are

[7] The canyons of southern California cut Miocene rocks in several places, and Monterey Canyon cuts a Pliocene formation.

not filled, processes must exist which are keeping them clean. The existence of mud flows is well demonstrated for the La Jolla canyons by the deepening which has been found on two occasions at the canyon heads. The small trench in the filled bottom of Sumner Branch (Pl. 3, fig. 2) is also particularly impressive. The La Jolla canyons have an abundance of kelp and eel grass within the fill, and this mixed with mud and sand should provide good sliding material in view of the steep bottom gradients.

Summary and Conclusions

Diving operations into the heads of the La Jolla submarine canyons have yielded new information regarding the details of these marine features. The telephonic descriptions during the course of the dives and the submarine photographs leave little doubt as to the exact character of the walls and the floors of the canyon. Scripps Canyon and its main tributary were found to be cut into firm Eocene conglomerate, sandstone, and limestone, and to have steep upper slopes and a vertical-walled inner gorge. Numerous gullies cut the vertical wall, some of them as hanging valleys and others as steeply descending accordant tributaries. Except for the thick cover of sediment on the steeply sloping canyon floor, Scripps Canyon appears to be remarkably similar to the canyons cut into the terrace and sea cliffs on the adjacent land.

The head of La Jolla Canyon is characterized by a network of tributaries which are excavated out of an alluvial formation similar to alluvium on the near-by land. These tributaries have high gradients and are characterized by their precipitous side walls. Some of the walls contain large masses of cobbles. Only one rock outcrop was encountered during the dives, although it is known that La Jolla Canyon has rock at somewhat greater depths. A hanging valley or a waterfall step was seen along the course of one tributary. The head of La Jolla Canyon has features which are almost exactly duplicated by the cuts in alluvium in the canyons north of Scripps Institution.

The floors of all the canyons had a cover of sediment. In the upper portions this sediment was sand, but beyond about 40- or 50-foot depths a considerable admixture of mud was encountered and decaying eel grass and kelp were abundant in most places. At the greatest depths there was less kelp, and the sediment was largely sand. In a few places cobbles were encountered on the valley floors.

The characteristics of these La Jolla marine canyons seem to leave little doubt of their stream origin. No submarine currents capable of excavation were encountered, nor were there ripples on the floors of the canyons except in a few places in very shallow water where wave oscillation was obviously responsible. New evidence was obtained that mud flows or landslides are operating to prevent the complete filling of the canyon heads. The hypothesis is proposed that the canyons were excavated at a relatively remote time and have undergone subsequent modifications in response to glacially controlled changes in sea level.

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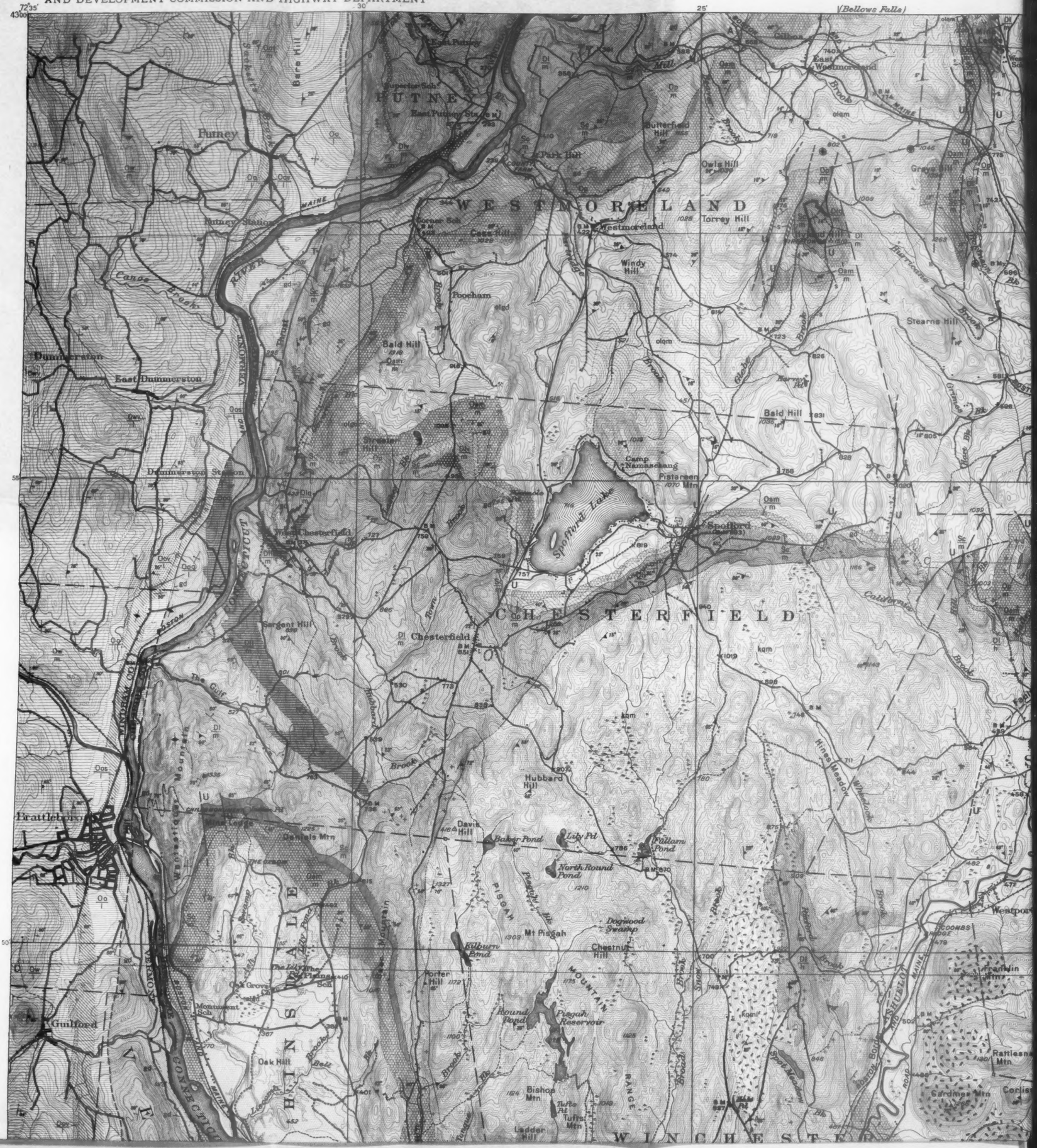
NEW HAMPSHIRE STATE PLANNING
AND DEVELOPMENT COMMISSION AND HIGHWAY DEPARTMENT



NEW HAMPSHIRE

KEENE AND BRATTLEBORO QUADRANGLES

LEGEND

METAMORPHIC AND IGNEOUS ROCKS

LATE DEVONIAN? — *New Hampshire Magma Series*

cg — Concord granite
(Light-gray fine-to medium-grained foliated granodiorite composed of oligoclase, quartz, and potash feldspar, with some muscovite and biotite.)

sqd — Spaulding quartz diorite
(Dark-gray medium-grained well-foliated granodiorite and quartz diorite composed of andesine, biotite, and quartz.)

kqmhi – kqm — Kinsman quartz monzonite
(Coarse-grained light-gray quartz monzonite to quartz diorite composed of quartz, oligoclase-andesine, potash feldspar, biotite, and muscovite; sillimanite present locally; in most places rock contains phenocrysts of microcline one to two inches long, locally phenocrysts are albite, oligoclase or andesine; border of body well foliated (kqm). Inclusions of mica schist and sillimanite schist abundant locally (kqmi).

gd — Granodiorite and quartz diorite
(Medium-grained light-to dark-gray granodiorite and quartz diorite composed of oligoclase-andesine, quartz, biotite, muscovite and potash feldspar. Locally foliated, some bodies strongly sheared.)

Amphibolite
(Areas where amphibolite sills and dikes are abundant. Dark-gray to black foliated amphibolite containing phenocrysts of andesine or labradorite; groundmass composed of hornblende and andesine or labradorite.)

MIDDLE DEVONIAN? — *Oliverian Magma Series*

olqd – olgd – olqm — Quartz diorite, granodiorite, and quartz monzonite gneiss
(Medium-grained gray to pink subporphyritic strongly -to weakly-foliated gneiss, composed of oligoclase-andesine, quartz, potash feldspar, and biotite. Depending on character and amount of feldspar, rock is quartz diorite (olqd), granodiorite (olgd) or quartz monzonite (olqm).)

LOWER DEVONIAN — *Lower Devonian*

Dl — Littleton formation
(Zone m: mica schist, quartz-mica schist, and thin fine-grained gray quartzite, locally with porphyroblasts of biotite, garnet, and/or staurolite (Dl). Local beds of white vitreous quartzite and quartz conglomerate (Dlq). Interbedded amphibolite and light-colored quartz-feldspar-biotite volcanics (Dlv). Interbedded lime-silicate granulite and schist, biotite schist, and fine-grained feldspathic quartzite, with some associated white vitreous quartzite, amphibolite, and light-colored quartz-feldspar-biotite volcanics(Dli) Zone h: mica schist, quartz-mica schist, and thin fine-grained gray quartzite, with porphyroblasts of garnet and sillimanite (Dl). White vitreous quartzite and quartz conglomerate (Dlq).

SILURIAN — *Lower or Middle*

Sc — Clough quartzite
(Zone m: coarse white vitreous quartzite and quartz conglomerate, with some thin beds of muscovite-quartz schist; porphyroblasts of garnet and/or staurolite present locally. Zone h: quartzite, quartz conglomerate, and some thin beds of muscovite-quartz schist. Porphyroblasts of garnet and traces of sillimanite present locally.)

ORDOVICIAN? — *Upper Ordovician?*

Op — Partridge formation
(Zone m: mica schist, garnet-mica schist, staurolite-mica schist, quartz-mica schist, and micaceous quartzite. Local lime silicate schist and granulite; also amphibolite and quartz-feldspar-biotite gneiss. Zone h: mica schist, garnet-mica schist, sillimanite-garnet schist, and micaceous quartzite.)

Oam — Ammonoosuc volcanics
(Zone m: chiefly fine-grained gray quartz-feldspar-biotite gneiss and dark amphibolite; occasional beds of white pyritiferous quartzite; local light-colored volcanic conglomerate; accessory garnet. Zone h: fine-grained

METAMORPHIC ZONES

Shown below formation symbol thus: Dl / m

l - low-grade
m - middle-grade
h - high-grade

CONTACTS

Accurate

Approximate and diagrammatic, due to poor exposure.

Indefinite as sharp contact is lacking.

Used to connect isolated outcrops of discontinuous stratigraphic units to indicate the inferred location of the horizon where the unit is absent.

SPECIAL SYMBOLS

(Strike and dip symbols represent only a small percentage of the field observations).

Strike and dip of bedding, including inverted and normal beds.

Strike and dip of inverted beds.

Strike of vertical beds.

Horizontal bedding.

Strike and dip of foliation and schistosity

Strike of vertical folintion and schistosity

Horizontal foliation and schistosity.

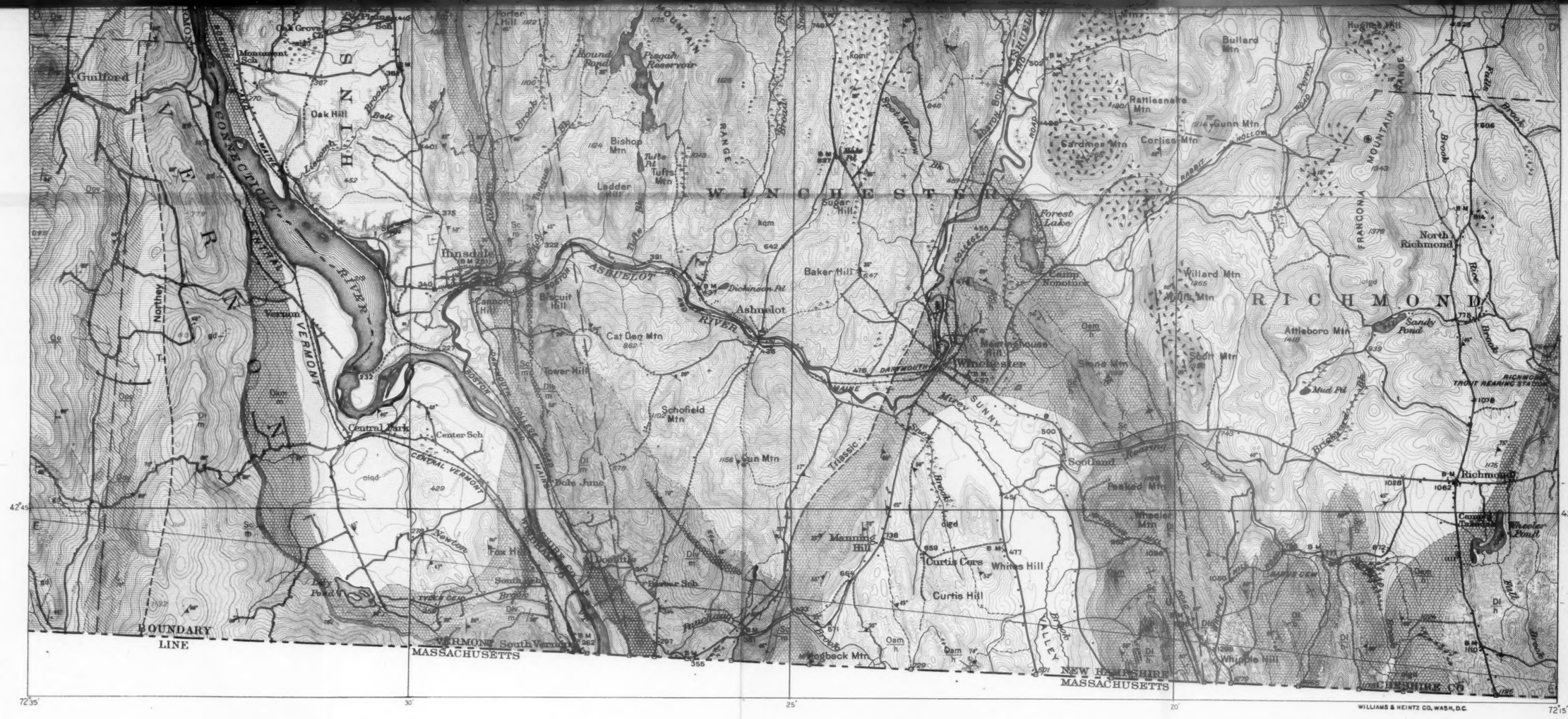
Normal fault; U on upthrown side (where space permits use of symbols).

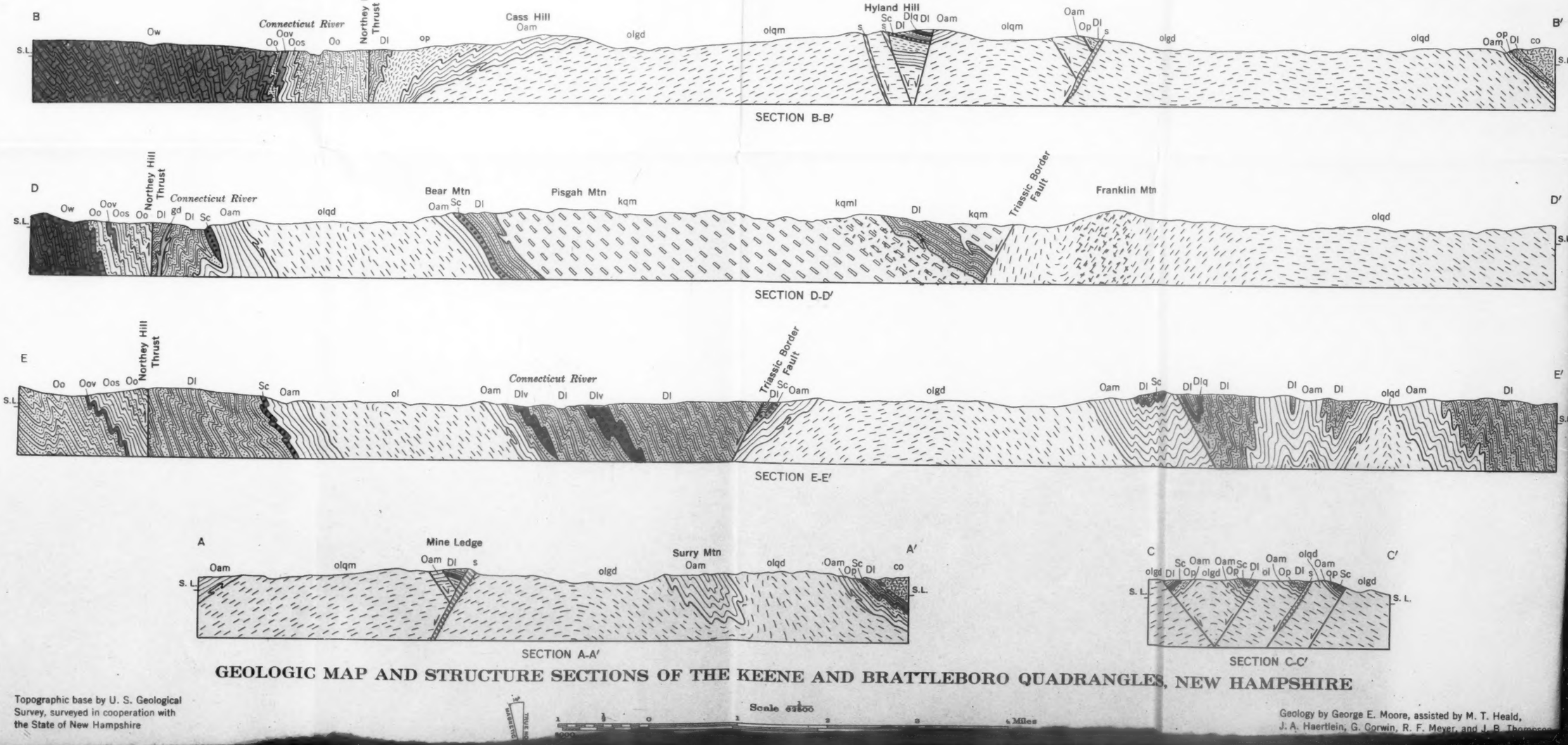
Thrust fault; T upthrown side.

Silicified fault zone.

Pegmatite

Mines and quarries, mostly abandoned.





GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE KEENE AND BRATTLEBORO QUADRANGLES, NEW HAMPSHIRE

Topographic base by U. S. Geological Survey, surveyed in cooperation with the State of New Hampshire

Scale 62500

1 0 1 2 3 4 Miles

1000 0 1000 2000 3000 4000 5000 6000 7000 8000 9000 10000 Feet

Geology by George E. Moore, assisted by M. T. Heald, J. A. Haertlein, G. Corwin, R. F. Meyer, and J. B. Thompson

WILLIAMS & HEINTZ CO., WASH., D.C.

SECTION B-B′

SECTION D-D′

SECTION E-E′

SECTION A-A′

SECTION C-C′

PRE-SILURIAN, PROBABLY ORDOVICIAN

Middle Ordovician? — *Upper Ordovician?*

Partridge formation

(Zone m: mica schist, garnet-mica schist, staurolite-mica schist, quartz-mica schist, and micaceous quartzite. Local lime silicate schist and granulite; also amphibolite and quartz-feldspar-biotite gneiss. Zone k: mica schist, garnet-mica schist, sillimanite-garnet schist, and micaceous quartzite.)

Ammonoosuc volcanics

(Zone m: chiefly fine-grained gray quartz-feldspar-biotite gneiss and dark amphibolite; occasional beds of white pyritiferous quartzite; local light-colored volcanic conglomerate; accessory garnet. Zone k: fine-grained quartz-feldspar-biotite gneiss, amphibolite, and some quartzite. Some of the quartzite and light-colored gneiss contain sillimanite.)

Lower part of Ammonoosuc volcanics, all of the Albee formation, and upper part of the Orfordville formation, which are present elsewhere, are missing in this quadrangle due to the Northey Hill thrust.

Orfordville formation

(Zone l: dark phyllite and slate, abundant thin gray quartzite and quartz-mica schist (Oo); fine grained buff to gray felsitic volcanics (Oov); quartzite, conglomerate with pebbles of quartz and slate, slate, and thin beds of impure marble (Ooa); dark slate with an occasional thin quartzite (Oos). Zone m: mica schist, quartz-mica schist, and quartzite, with porphyroblasts of biotite, garnet, and or staurolite.)

Waits River formation

(Zone l: fine-grained dark-gray impure marble, dark slate, and phyllite, with interbedded thin quartzite and quartz-mica schist (Ow): light-gray fine-grained felsitic volcanics (Owv). Zone m: fine-grained dark-gray impure marble, mica schist, quartz-mica schist, and thin quartzite, with porphyroblasts of biotite or garnet (Ow); amphibolite with some quartz-feldspar-biotite gneiss (Owv).

Mines and quarries, mostly abandoned.

GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE KEENE AND BRATTLEBORO QUADRANGLES, NEW HAMPSHIRE

Scale 1/62500

1 0 1 2 3 4 Miles

5000 0 5000 10000 15000 20000 Feet

1 0 1 2 3 4 Kilometers

Contour interval 20 feet

Datum is mean sea level

Geology by George E. Moore, assisted by M. T. Heald, J. A. Haertlein, G. Corwin, R. F. Meyer, and J. B. Thompson. Directed by Marland P. Billings. Geology surveyed in 1940, 1941, 1942, and 1946, with the aid of grants from the Whitney, Sayles, and Holden funds of Harvard University. Published 1948.

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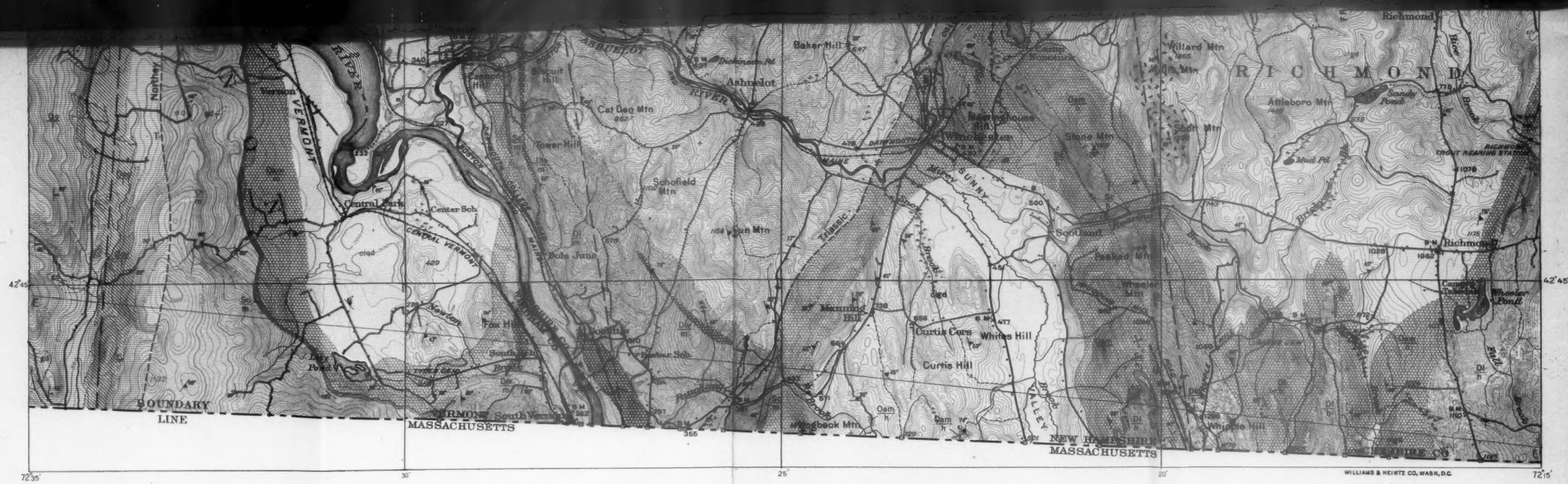

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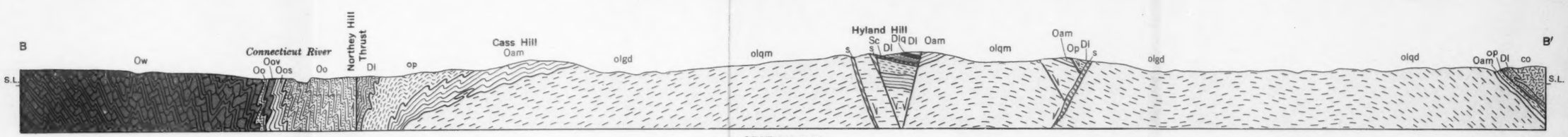
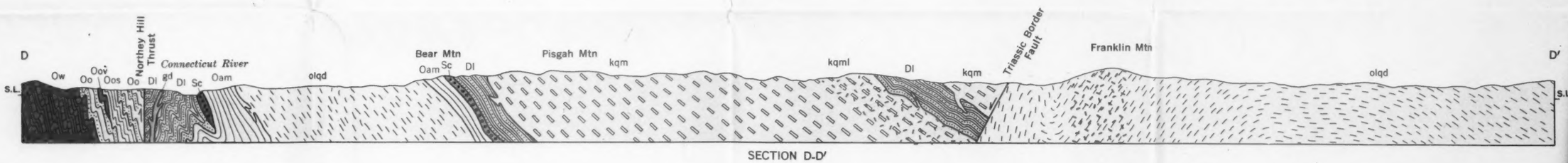


SECTION B-B'



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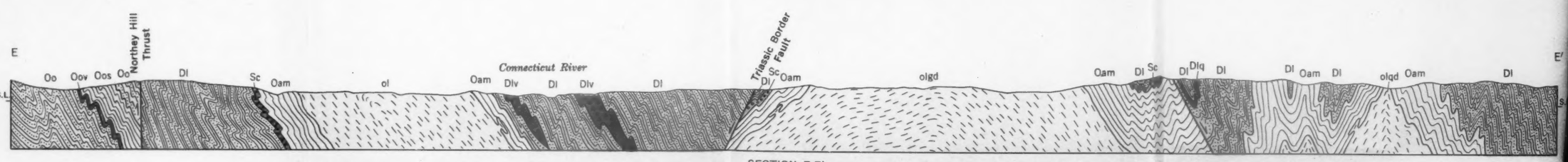

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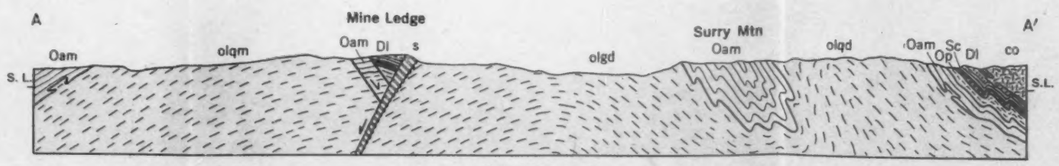

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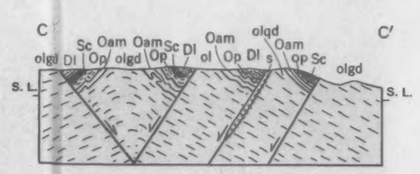
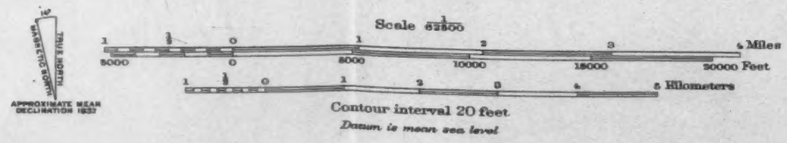


SECTION C-C'

GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE KEENE AND BRATTLEBORO QUADRANGLES, NEW HAMPSHIRE

Topographic base by U. S. Geological Survey, surveyed in cooperation with the State of New Hampshire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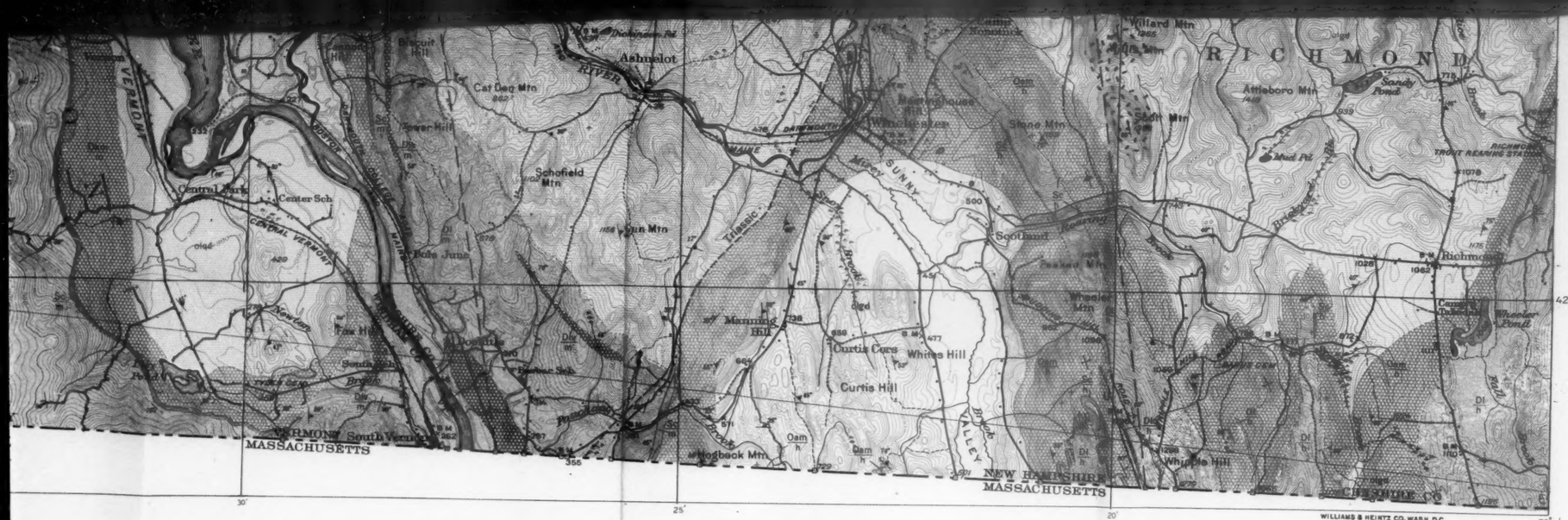
5000 0 5000 10000 15000 20000 Feet
1 0 1 2 3 4 5 Kilometers
Contour interval 20 feet
Datum is mean sea level

Geology by George E. Moore, assisted by M. T. Heald, J. A. Haertlein, G. Corwin, R. F. Meyer, and J. B. Thompso
Directed by Marland P. Billings. Geology surveyed in 1940
1941, 1942, and 1946, with the aid of grants from the Whitney, Sayles, and Holden funds of Harvard University.
Published 1948.



PRE-SILURIAN, PROBABLY Middle Ordovician?

Oo
Oos
Oov
Oo

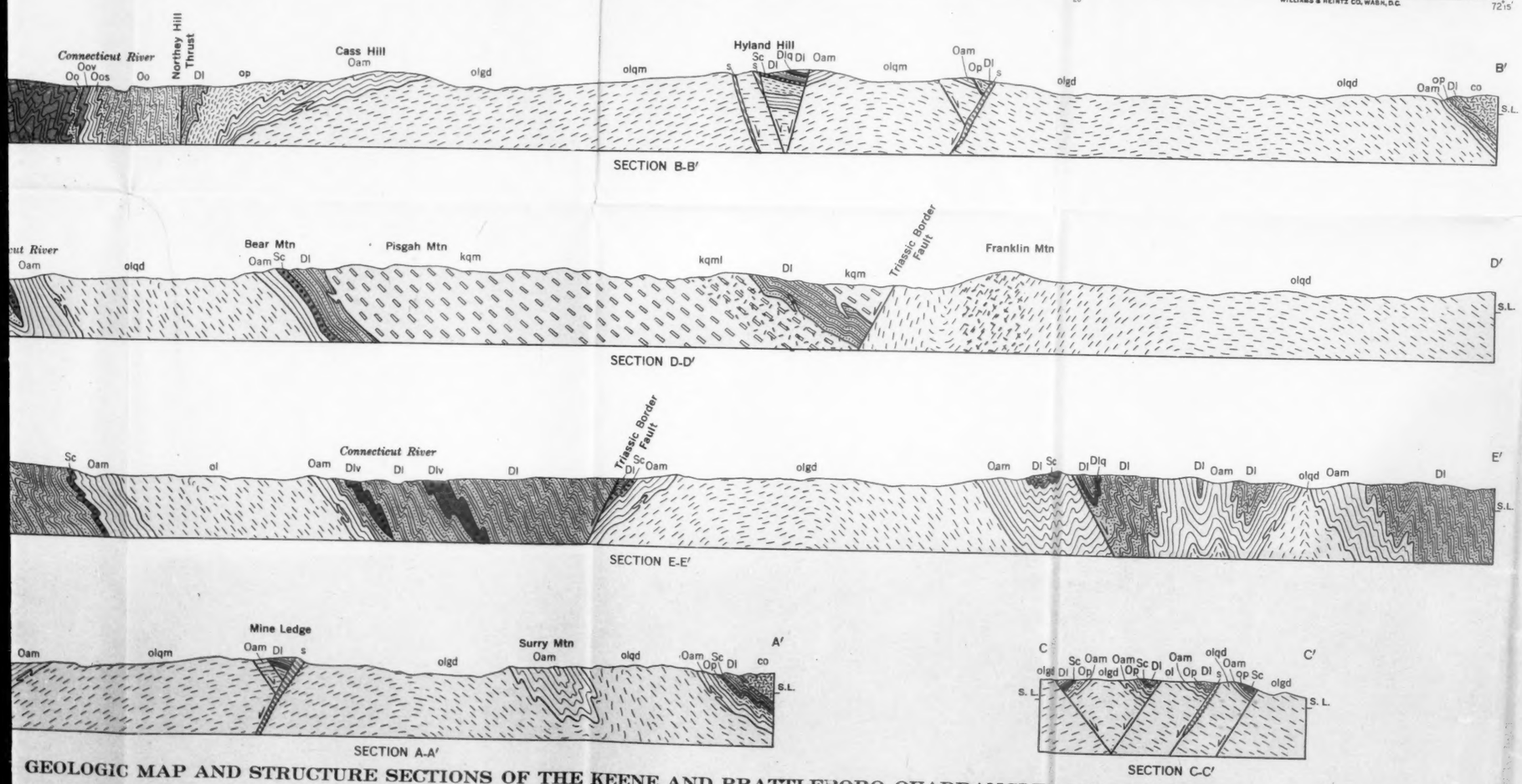
Orfordville formation

(Zone l: dark phyllite and slate, abundant thin gray quartzite and quartz-mica schist (Oo); fine-grained buff to gray felsitic volcanics (Oov); quartzite, conglomerate with pebbles of quartz and slate, slate, and thin beds of impure marble (Oos); dark slate with an occasional thin quartzite (Oos). Zone m: mica schist, quartz-mica schist, and quartzite, with porphyroblasts of biotite, garnet, and or staurolite.)

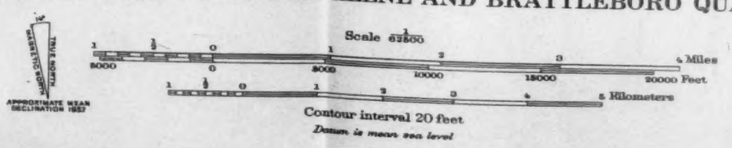
Owm
Ow

Waits River formation

(Zone l: fine-grained dark-gray impure marble, dark slate, and phyllite, with interbedded thin quartzite and quartz-mica schist (Ow); light-gray fine-grained felsitic volcanics (Own). Zone m: fine-grained dark-gray impure marble, mica schist, quartz-mica schist, and thin quartzite, with porphyroblasts of biotite or garnet (Ow). amphibolite with some quartz-feldspar-biotite gneiss (Owm).



GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE KEENE AND BRATTLEBORO QUADRANGLES, NEW HAMPSHIRE



Geology by George E. Moore, assisted by M. T. Heald, J. A. Haertlein, G. Corwin, R. F. Meyer, and J. B. Thompson. Directed by Marland P. Billings. Geology surveyed in 1940, 1941, 1942, and 1946, with the aid of grants from the Whitney, Sayles, and Holden funds of Harvard University. Published 1948.

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 60, PP. 1613-1670, 5 FIGS., 9 PLS. OCTOBER 1949

STRUCTURE AND METAMORPHISM OF THE KEENE-BRATTLEBORO AREA, NEW HAMPSHIRE-VERMONT

By George E. Moore, Jr.

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Abstract

Metamorphosed sedimentary and volcanic rocks, ranging in age from middle Ordovician (?) to lower Devonian, occupy about 55 per cent of the Keene-Brattleboro area of southwestern New Hampshire and southeastern Vermont. Rocks of the low-grade zone of metamorphism are present in a north-south band near the western edge of the area; rocks of the middle-grade zone occur on either side of the low-grade zone; and rocks of the high-grade zone occupy the central and eastern parts of the area.

The lithology of the stratified rocks is similar to that farther north. The Bernardston formation of north-central Massachusetts is correlated with the Littleton formation.

Foliated igneous rocks of middle Devonian (?) and late Devonian (?) age occupy about 45 per cent of the area. These include three plutons of the Oliverian magma series and five members of the New Hampshire magma series.

The stratified rocks were severely folded, broken by a major thrust fault, and metamorphosed during the late Devonian (Acadian) orogeny. The resulting structural features in this area, trend almost north. The Bronson Hill anticline occupies most of the area. Three elongate domes, each with a plutonic core of the Oliverian magma series and a surrounding envelope of stratified rock, are present on this anticline. A pluton of Kinsman quartz monzonite occurs in a structural basin on the anticline, near the center of the area. The steeply dipping western limb of the Brattleboro syncline occupies the western one-eighth of the area. The Northey Hill thrust, a major fault of western New Hampshire, separates the Bronson Hill anticline from the Brattleboro syncline. This thrust, which cuts out the eastern limb of the Brattleboro syncline, has a stratigraphic separation of about 10,000 feet.

Foliation and bedding are mutually parallel in most of the stratified rocks east of the Northey Hill thrust, whereas foliation is parallel to the axial planes of folds in most of the rocks west of the thrust. Minor folds of two ages are present west of the thrust and in a few places just east of thrust. The older minor folds are isoclinal, or nearly so, and have a well-developed axial-plane flow cleavage. The younger minor folds are more open and have an incipient to strong axial-plane slip cleavage.

It is believed that the Oliverian magma series

was forcefully injected as a thick composite sheet during the very early stages of the orogeny, while the enclosing rocks were essentially flat. The New Hampshire magma series was emplaced during the late stages of the orogeny. Initial movements along the Northey Hill thrust may have occurred during early stages of the orogeny, but a considerable part of the movement took place after the metamorphic intensity in the rocks along the fault had reached its maximum.

The Triassic border fault, which forms the eastern border of the Triassic rocks of the Connecticut valley in Massachusetts, enters the area 1 mile east of the Connecticut River and trends northeasterly to the vicinity of Keene, where it apparently ends. The maximum stratigraphic displacement along the fault is about 15,000 feet. Numerous small high-angle normal faults are present. In three localities, graben, in which the roof of one of the plutons of the Oliverian magma series is preserved, form very striking structural features. The normal faults which bound the graben have a maximum stratigraphic separation of 3000 to 4000 feet. The age of the faulting associated with the graben, and also much of the normal faulting of western New Hampshire, is believed to be Triassic. Fluorite deposits of the Keene-Brattleboro area are also believed to be Triassic.

Introduction

The Keene-Brattleboro area covers about 315 square miles in the extreme southwestern corner of New Hampshire and the southeastern corner of Vermont (Fig. 1). It includes the Keene quadrangle (1935), the eastern third of the Brattleboro quadrangle (1935), and the northeastern corner of the fifteen-minute Greenfield quadrangle, Massachusetts-Vermont (1894).

The Keene-Brattleboro area is maturely dissected. Although the maximum relief is about 1300 feet, local relief in most places is only a few hundred feet. The Connecticut River, the west bank of which is the boundary between New Hampshire and Vermont, flows south across the western part of the area. Most of the region is forested; the larger valleys and some of the gentle higher slopes are tilled. Outcrops are scattered, but bedrock is not completely covered over any large area.

Early studies of the geology of the area include those of C. T. Jackson (1844), E. Hitchcock (1861), C. H. Hitchcock (1874; 1877; 1878), and R. A. Daly (1897). C. H. Richardson (1933) did more recent work in the Vermont portion of the area. In recent years a great deal of detailed work has been done in New Hampshire, Vermont, and Massachusetts, much of which has been of great value in the present study. The work of M. P. Billings (1937) in the Littleton-Moosilauke area (Fig. 1) has proved to be of particular importance.

Three and one-half summers, between 1940 and 1946, were spent in mapping the geology of the area. United States Geological Survey topographic sheets, enlarged to a scale of 3 inches to 1 mile, were used as base maps. Outcrops were located by inspection, by pace and compass traverse, or by means of an aneroid barometer. Detailed pace and compass maps, on a scale of 500 feet to an inch, were made in areas where the structure was relatively complex or where it was desired to trace very accurately the contacts of formations. Numerous hand specimens were collected during the field work, from which 125 thin sections were prepared and studied.

Acknowledgments

The aid and counsel of Professor Marland P. Billings of Harvard University, under whose guidance this work was undertaken, is sincerely appreciated. Professor Billings spent several weeks in the area each summer the work was in progress and contributed a great deal of help in the preparation of the manuscript. Professor E. S. Larsen, Jr., gave valuable aid both in the study of thin sections and in the preparation of the manuscript.

Grateful acknowledgment is made to the Sayles Fund and the Whitney Fund of Harvard University, whose generous grants greatly aided in financing the field work, and to the Department of Mineralogy and Petrography, Harvard University, who paid for most of the thin sections.

The writer was ably assisted in the field work by Milton T. Heald and James A. Haertlein during the summer of 1941, by Gilbert Corwin during the summer of 1942, and by James B. Thompson and Richard F. Meyer during the summer of 1946.

Discussion of common problems with members of adjoining field parties, particularly those of Dr. F. C. Kruger (Bellows Falls quadrangle, in 1940), M. T. Heald (Lovewell Mountain quadrangle, in 1946), Dr. Katharine Fow-

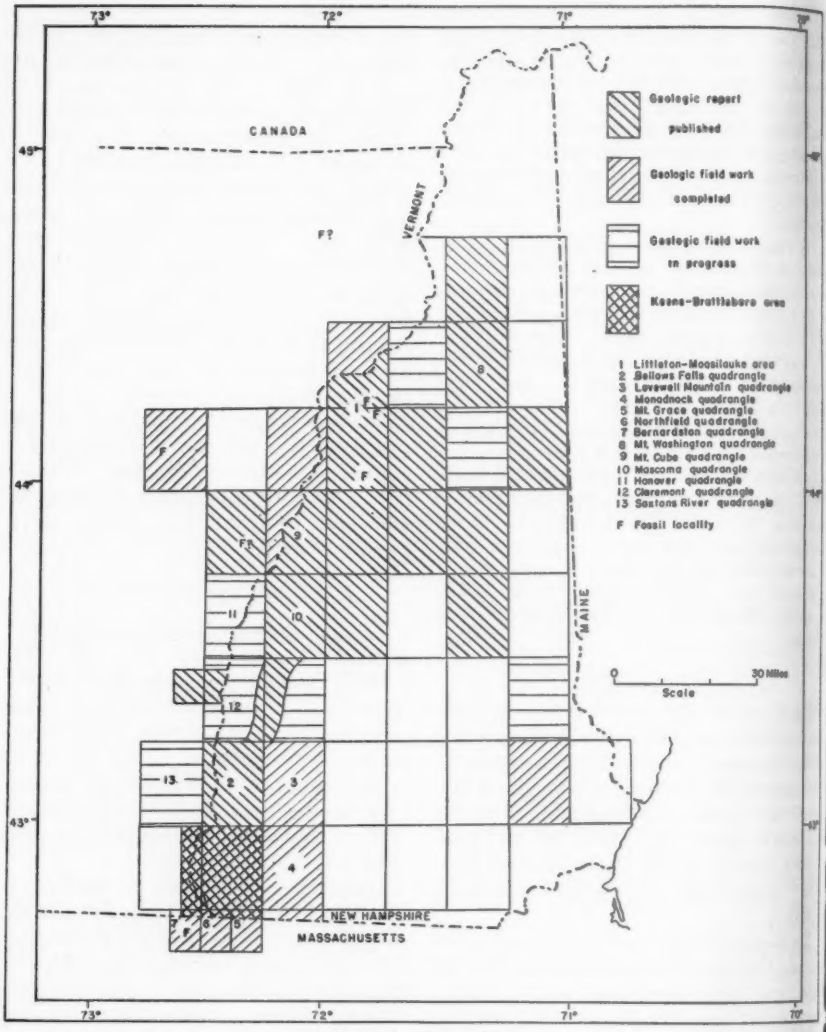

FIGURE 1.—INDEX MAP

ler-Billings (Monadnock quadrangle, in 1946), Dr. J. B. Hadley (Mt. Grace quadrangle, in 1946), and Dr. Robert Balk (Northfield and Bernardston quadrangles, in 1946), was of great value.

GENERAL LITHOLOGIC FEATURES OF THE STRATIFIED ROCKS

Metamorphosed sedimentary and volcanic rocks occupy about 55 per cent of the Keene-Brattleboro area. The original sedimentary rocks were conglomerates, sandstones, shales, and calcareous beds. The volcanic rocks include both rhyolitic and basaltic types. These stratified rocks fall into six major stratigraphic units ranging in age from middle Ordovician (?) to lower Devonian (Fig. 2). As noted in the columnar section, the Albee formation, exposed elsewhere, is cut out in the Keene-Brattleboro area by a major thrust fault.

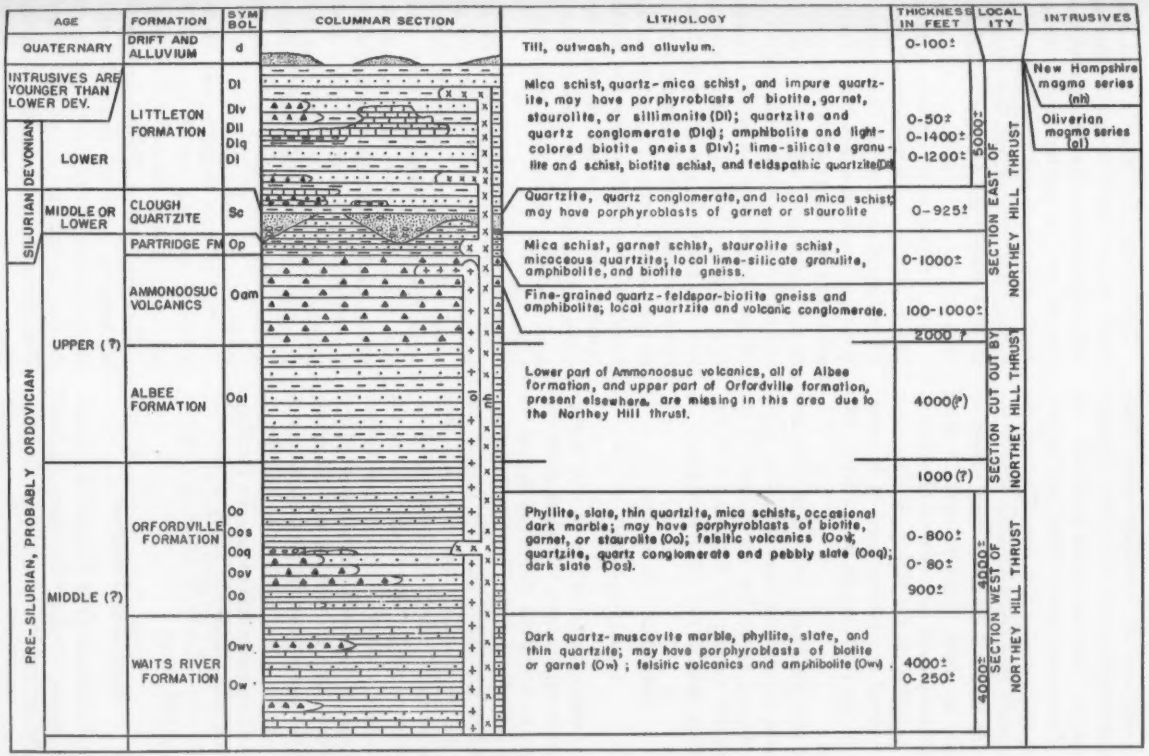

FIGURE 2.—COLUMNAR SECTION FOR KEENE-BRATTLEBORO AREA

The grade of metamorphism ranges from low-grade to high-grade (Pl. 4). The low-grade metamorphic rocks occur in a band just west of the Connecticut River. Albite, sericite, and chlorite are typical of this zone. Rocks of the middle-grade zone occur on either side of the low-grade zone and contain biotite, garnet, or staurolite as index minerals; some of these minerals are present as coarse porphyroblasts. Rocks belonging to the high-grade zone occur in the central and eastern parts of the area. Many of the rocks in this zone contain sillimanite.

No fossils were found. Correlations were made by tracing recognizable formations from previously mapped areas. The major units of the stratigraphic column are very similar to corresponding units at their type localities.

Ordovician (?) Strata

General Statement

The Ordovician (?) strata of the Keene-Brattleboro area include the Waits River formation, the Orfordville formation, the Ammonoosuc volcanics, and the Partridge formation. The Albee formation, normally present beneath the Ammonoosuc volcanics, has been cut out by the Northey Hill thrust, which extends from a point about 4 miles east of the northwestern corner of the area to a point about 1 mile east of the southwestern corner of the area (Pl. 1). The Waits River and Orfordville formations are present in north-south bands on the western side of the Northey Hill thrust. The Ammonoosuc volcanics and the Partridge formation form concentric bands around the plutonic cores of several domes east of the Northey Hill thrust.

Waits River Formation

General statement.—The Waits River formation crops out in a general north-south band along the extreme western edge of the area. The eastern part of the belt is in the low-grade zone of metamorphism, the western part is in the middle-grade zone (Pl. 4).

Low-grade zone.—The Waits River formation, in the low-grade zone, is composed of phyllite and slate with interbedded quartz schist and quartzite, arenaceous marble, and some felsitic volcanics.

Phyllite and slate are the most abundant rocks. Most of the dark-gray to black slate is too crinkled for use as a roofing slate; however, the rock has been quarried in one or two places. Much of the slate and phyllite has both flow cleavage and slip cleavage. The essential minerals of these rocks are quartz, sericite, and chlorite.

Beds of dark-gray fine-grained quartz-muscovite marble are characteristic of the Waits River formation (Pl. 5, fig. 1). Leaching of the carbonate produces a dark-brown fine-grained soft punky rock composed of the insoluble part of the marble beds. In places protected from rapid erosion, this punky layer may be several inches thick. In the more westerly exposures of this formation, beds of impure marble constitute as much as 50 per cent of the formation, and individual beds have a maximum thickness of several tens of feet. The beds of impure marble decrease in abundance and in thickness easterly toward the top of the formation, where they make up only a few per cent of the total thickness, and none is more than a few feet thick.

Quartz-sericite schist and quartzite are interbedded with the slate, phyllite, and impure marble. Beds of these light-gray fine-grained rocks are also thicker and more abundant toward the west, where some of them are 20-30 feet thick. They thin eastward toward the top of the formation, where most of them are a fraction of an inch to an inch or so thick. In the upper part of the formation the proportion of these thin beds differs greatly from place to place.

A lens of gray to gray-green fine-grained volcanics, near the top of the Waits River formation, extends southward from the village of East Dummerston. The lens has a maximum thickness of about 400 feet and a length of about 10,000 feet. In some places these rocks show a banding that appears to be bedding. The volcanics contain grains of plagioclase, quartz, and biotite, 0.5 to 1.0 millimeter in diameter, in a very fine-grained groundmass of quartz and plagioclase (Table 1, column 1). Although the specimen listed in Table 1 contains some biotite, an index mineral of the middle-grade zone, the surrounding rocks are typical low-grade rocks.

Middle-grade zone.—The more westerly ex-

TABLE 1.—ESTIMATED MODES OF THE WAITS RIVER FORMATION

	Low-grade zone	Middle-grade zone				
	1	2	3	4	5	6
Relic phenocrysts						
Quartz	10	—	—	—	—	—
Porphyroblasts						
Plagioclase	6	—	—	—	—	—
Biotite	2	—	—	—	—	—
Chlorite	3	—	—	—	—	—
Groundmass						
Quartz	72	45	35	52	tr	33
Plagioclase		—	—	tr	tr	tr
Hornblende	—	—	—	—	75	7
Sericite	4	—	—	—	—	—
Muscovite	—	8	15	25	—	—
Biotite	—	—	—	22	—	—
Calcite	—	45	48	—	tr	5
Magnetite-ilmenite	—	tr	tr	—	tr	tr
Leucoxene	1	—	—	—	—	—
Hematite	—	1	—	—	—	—
Pyrite	tr	1	2	1	—	—
Tourmaline	—	—	—	tr	—	—
Zircon	—	—	tr	tr	—	—
Sphene	—	—	—	—	tr	—
Apatite	tr	—	—	tr	—	—
Rutile	—	—	—	—	tr	—
Epidote	2	—	—	—	25	55
Per cent of anorthite in plagioclase	25*	—	—	—	—	—
Size of phenocrysts or porphyroblasts (mm)	0.5–1.0	—	—	—	—	—
Size of groundmass (mm)	0.025	0.1–0.2	0.07–0.15	0.2–0.3	0.3–1.0	0.3–0.5
Texture	S″	Gr′	S	S	S	S

*Probably a relic mineral.
″S—Schistose.
′Gr—Granoblastic.

1. Felsitic volcanic.
2. Quartz-muscovite marble.
3. Quartz-muscovite marble.
4. Biotite schist.
5. Amphibolite.
6. Hornblende-epidote gneiss.

posures of the Waits River formation in this area are in the middle-grade zone. Mica schist, quartz schist, quartzite, and impure marble, with porphyroblasts of biotite or garnet, constitute the formation. Westward from the low-grade zone, biotite appears first in the impure quartzites and quartz schists. Biotite is the more common porphyroblast in these middle-grade rocks, but garnet is abundant in some places. A small body of amphibolite and light-colored gneiss, present at the western edge of the area near the village of Dummerston, is probably metamorphosed tuff. Table 1 gives the mineral composition and estimate modes of some rocks of the middle-grade zone.

Thickness.—Repetition of beds by folding makes it impossible to determine accurately the thickness of exposed strata. A fair estimate is 4000 to 5000 feet. The base of the formation is not exposed.

Orfordville Formation

General statement.—The Orfordville formation crops out in a north-south belt between the Northey Hill thrust on the east and the Waits River formation on the west. The mineralogy of most of the rocks of the southern half of this belt is characteristic of the low-grade zone (Pl. 4). The intensity of metamorphism increases toward the north, and, at the northern edge of the area, staurolite porphyroblasts are common in the eastern (upper) half of the formation. In some places near the Northey Hill thrust (Pl. 4), garnet porphyroblasts have been completely altered to sericite or chlorite during retrograde metamorphism.

The Orfordville formation contains several distinctive members which, in a general way, can be traced parallel to the strike (Pl. 1). The lowest part of the formation consists of phyllite with many interbedded thin quartz-sericite schists and quartzites. Above this is a discontinuous band of felsitic volcanics. Overlying the volcanics are slates, phyllites, or mica schists, which, in a few places, have a quartz conglomerate near their base. The uppermost member of the formation consists of phyllite or mica schist, interbedded with thin quartzites and quartz-sericite schists. A few thin beds of quartz-muscovite marble are present in the lower half of the formation.

Low-grade zone: lower phyllite-quartzite member.—The lowest part of the Orfordville formation is dark-gray fine-grained phyllite (Pl. 6, fig. 1) containing many beds of medium-gray fine-grained quartz-sericite schist and quartzite, ½ to 2 inches thick, and a few thin lenses of fine-grained quartz-muscovite marble. Nearly every outcrop of this member shows numerous small-scale folds, many of them doubly plunging. The member is very similar to, and may be the equivalent of, part of the Gile Mountain formation (Doll, 1944), the type locality for which is in the Strafford quadrangle, 75 miles to the north.

The phyllite is composed of sericite, chlorite, quartz, and albite, with a grain size of 0.05 to 0.1 millimeter. The quartz-sericite schists and quartzites contain the same minerals in different proportions. These rocks were originally dark shales and impure sandstones. An estimated mode of phyllite is given in Table 2, column 1.

The phyllite-quartzite member is 1000 to 1500 feet thick.

Low-grade zone: volcanic member.—The discontinuous lenses of volcanics in the Orfordville formation are very fine-grained buff or gray rocks with a greasy luster; on weathered surfaces they are gray green and have a silky luster. The volcanics show little bedding except toward the top and bottom of the member, where they contain admixed sedimentary material. Cleavage is best developed in the beds of mixed origin. Fragmental texture is present in a few places. Microscopic examination reveals relic phenocrysts of quartz 0.04 to 0.1 millimeter in diameter. Porphyroblasts of oligoclase and muscovite of similar size are also present. As shown in Table 2, columns 2, 3, and 4, the groundmass is quartz, chlorite, and sericite with accessory leucoxene, pyrite, carbonate, tourmaline, apatite, and zircon in grains averaging 0.005 to 0.02 millimeter in diameter.

These rocks, in the field, have the aspect of felsites, and because of their chemistry, inferred from their mineralogy, they are considered to have been rhyolitic. It is apparent, however, that they have been extensively metamorphosed. Although most of these volcanics were probably deposited as water-laid tuffs, some may have been flows. The maximum thickness of the volcanic member is about 800 feet.

Low-grade zone: slate member.—This member is a rather uniform dark-gray slightly quartzose and pyritiferous rock, which in places contains a few thin beds of quartz-mica schist or quartzite. The slaty cleavage is an axial-plane foliation that cuts across bedding on the noses of folds, but on the long limbs of the folds the cleavage appears to parallel the bedding. Slate has been quarried commercially at six or more localities along the strike (Pl. 1). In most places, however, the rock contains crinkles with an axial-plane slip cleavage which make the rock unsuitable for industrial purposes. The maximum thickness of the slate member is about 900 feet.

A discontinuous intraformational conglomerate with a maximum thickness of about 80 feet is present near the base of the slate member about 4 miles north of the city of Brattleboro

TABLE 2.—ESTIMATED MODES OF THE ORFORDVILLE FORMATION

	Low-grade zone				Middle-grade zone		
	1	2	3	4	5	6	7
Relic phenocrysts							
Quartz	—	16	5	—	—	—	—
Porphyroblasts							
Plagioclase	—	10	2	10	—	—	—
Biotite	—	—	—	—	—	—	8
Muscovite	—	10	—	—	—	—	—
Chloritized garnet	—	—	—	—	4	3	—
Leucoxene	—	—	—	—	2	2	—
Limonite	5	—	—	—	—	—	—
Groundmass							
Quartz	35	18	33	35	30	35	46
Plagioclase		—	—	—	—	—	—
Biotite	—	—	—	—	—	tr	—
Chlorite	5	10	—	—	16	20	—
Sericite	55	34	35	45	48	40	46
Calcite	—	tr	25	10	—	—	tr
Epidote	—	—	—	—	—	tr	—
Tourmaline	—	tr	—	—	tr	tr	—
Apatite	—	tr	tr	—	—	—	—
Zircon	—	tr	tr	—	tr	—	tr
Rutile	—	—	—	—	—	tr	—
Leucoxene	—	2	—	tr	tr	—	—
Magnetite	—	—	—	—	—	—	tr
Pyrite	tr	tr	—	—	—	—	tr
Hematite	—	—	—	—	tr	—	—
Per cent of anorthite in plagioclase	—	13	12	10	—	—	—
Size of phenocrysts or porphyroblasts (mm)	1.0–2.0	0.04–0.08	0.05–0.1	0.05–0.1	0.5–1.5	0.5–1.0	0.3–0.5
Size of groundmass (mm)	0.05–0.1	0.005–0.01	0.01–0.02	0.01–0.02	0.15–0.8	0.05–0.1	0.005–0.01
Texture	S*	S	S	S	S	S	Gr**

*S—Schistose.
**Gr—Granoblastic.

1. Crinkled phyllite.
2. Felsitic volcanic.
3. Felsitic volcanic.
4. Felsitic volcanic.
5. Chloritized garnet schist.
6. Chloritized garnet schist.
7. Fine-grained black sericitic quartzite.

The lower 55 feet consists of slate, quartzose slate, and interbedded quartzites. The quartzites, in beds as much as several inches thick, are light-colored fine-grained rocks composed almost entirely of quartz. The upper 25 feet is quartzite and conglomerate with a few thin beds of dark-gray quartzose marble. Some of the conglomerates are composed of quartz pebbles as much as 6 or 8 inches in diameter, in a light-colored quartzite or dark slaty matrix. Other beds contain slate pebbles up to 1½ feet long and 2 or 3 inches thick, in a similar matrix. The conglomerate thins rapidly along its strike. Three miles north of the city of Brattleboro, the conglomerate is represented by 25 or 30 feet of dark slate containing 5–10 per cent of

quartz pebbles. The conglomerate may be the equivalent of the Hardy Hill quartzite, which is present about 20 miles to the north in the Bellows Falls quadrangle (Kruger, 1946, p. 168).

Low-grade zone: upper phyllite-quartzite member.—This member consists of phyllite and slate with interbedded quartz-sericite schist and quartzite (Pl. 5, fig. 2). Thin quartzites are so numerous in parts of this member that it acquired the field name of "ribbon rock". Individual beds are so similar to those of the lower member of the Orfordville formation that further description is not necessary. The member is 1000 to 2000 feet thick.

Middle-grade zone.—The middle-grade rocks of the Orfordville formation of this area are confined to a belt immediately west of the Northey Hill fault (Pl. 4). The maximum width of the belt, 1¼ miles, is near the northern edge of the quadrangle. Biotite, garnet, and/or staurolite are present in the rocks of this belt. These minerals are more abundant in, but not confined to, the beds equivalent to the phyllite and slate of the low-grade zone. Garnet reaches a maximum diameter of several millimeters, staurolite a length of 3 to 5 centimeters. The groundmass of the rocks in the middle-grade zone is considerably coarser than that of equivalent low-grade rocks.

Numerous beds of fine-grained black marble, black argillite, and fine-grained black quartzite, rock types not found in the Orfordville formation elsewhere in the area, are exposed along a brook 0.8 mile east of the village of East Dummerston. Individual beds have a maximum thickness of about 1 foot. These black beds contain pyrite and graphite and show very little foliation.

Modes of some of the rocks from the middle-grade zone are given in Table 2.

Thickness.—The complexity of folding in the Orfordville formation prohibits an accurate determination of its thickness. A fair estimate of the maximum thickness in this area is about 4000 feet. The upper part of the formation has been cut out by the Northey Hill thrust.

Missing Formations

The Albee formation (Billings, 1935, p. 9), present further northeast in New Hampshire (Billings, 1937; Hadley, 1942), is absent in the Keene-Brattleboro area because of thrust faulting. The formation is included in the columnar section (Fig. 2) and is mentioned here for completeness. The Albee formation overlies the Orfordville formation and underlies the Ammonoosuc volcanics; in the middle-grade zone of metamorphism it consists of quartzite, quartz-mica schist, and mica schist (Hadley, 1942, p. 124).

Ammonoosuc Volcanics

General statement.—The Ammonoosuc volcanics, composed of light-colored gneiss and dark amphibolite, occur in continuous bands, of varying width, adjacent to the rocks of the Oliverian magma series. Along with younger formations, these volcanics are also present in graben in the northeastern part of the area. The Ammonoosuc is present in both the middle-grade and high-grade zones of metamorphism.

Bedding is well defined throughout most of the formation. The beds range in thickness from a fraction of an inch to several feet. Foliation, expressed by both platy and needle-shaped minerals, is also well developed and is rigorously parallel to the bedding. Many of the light-colored volcanics contain slightly flattened quartz and feldspar grains whose intermediate and long axes lie in the plane of the foliation.

Most of the volcanics are metamorphosed water-laid crystal tuffs and breccias. Well-rounded pebbles are locally present; original fragmental texture is preserved in a few places. More massive light-colored beds, found near the top of the formation in some places, may have been flows. A few thin beds of quartzite and schist are interbedded with the volcanics.

Middle-grade zone.—The rocks of the Ammonoosuc formation west of a line between the village of Winchester and the western side of Surry Mountain are in the middle-grade zone of metamorphism (Pls. 1, 4). The dominant rocks are light-colored biotite-quartz-feldspar gneiss (Pl. 6, fig. 3) and dark amphibolite; light-colored actinolite-quartz-feldspar gneiss is not uncommon. These rocks are interbedded. Some of the light-colored gneisses, especially those in the southwestern part of the area, contain numerous relic phenocrysts of milky or blue quartz. In a few places the gneiss contains pebbles of quartz-

ite, or pebbles of light-colored biotite-quartz-feldspar gneiss similar in composition to that of the matrix. The pebbles are flattened, range in size up to 1 by 4 by 5 inches, and in places produce a strong lineation. The light-colored gneisses are somewhat more abundant than the amphibolites.

The light-colored biotite-quartz-feldspar gneiss is a fine-grained sugary-textured rock whose essential minerals are plagioclase, quartz, and biotite. The plagioclase, in different specimens, ranges from An_9 to An_{30}. Relic phenocrysts of quartz, as much as 15 millimeters in diameter, and small porphyroblasts of garnet are locally abundant. Chlorite, muscovite, garnet, and staurolite are included in a long list of accessory minerals. The groundmass of these rocks has an average grain size of 0.15 to 0.3 millimeter. Most of these light-colored volcanics were originally dacites. Estimated modes are given in Table 3, column 4. Although the average rock of the Ammonoosuc volcanics contains little or no potash feldspar, a light-colored quartz-feldspar gneiss from an elevation of 400 feet on Ox Brook in Westmoreland township contains 15 per cent orthoclase (Table 3, column 3).

Some of the light-colored beds of the Ammonoosuc volcanics differ from the more abundant biotite-quartz-feldspar gneiss in that they contain an appreciable amount of actinolite or hornblende, either as porphyroblasts or in the groundmass (Table 3, columns 5, 6).

The amphibolite, medium-grained well-foliated dark-gray to almost black, is composed of hornblende and plagioclase with varying amounts of accessory quartz, magnetite, garnet, biotite, potash feldspar, apatite, and sphene. Most of the amphibolite contains some chlorite. The hornblende of these rocks, the common blue-green variety, occurs either as needle-shaped or blade-shaped grains. Some of the amphibolite in which the hornblende occurs as needles shows prominent lineation. The hornblende content of these rocks is between 38 and 70 per cent, and they contain between 20 and 50 per cent plagioclase, the composition of which, in different specimens, ranges from An_{20} to An_{40}. The average grain size of the hornblende is between 1.5 and 2.0 millimeters; that of the feldspar is about 0.25 millimeter. The amphibolite is the metamorphosed equivalent of basalt or andesite. Table 3, column 7 gives an average mode of the amphibolite.

Occasional thin beds of fine-grained very light-colored muscovite-pyrite quartzite (Table 3, column 1), sericite schist (Table 3, column 2), and quartz-mica schist are present in the volcanics.

High-grade zone.—The rocks of the Ammonoosuc volcanics east of a line between the village of Winchester and the western side of Surry Mountain are in the high-grade zone of metamorphism. The more common rocks are light-colored biotite-quartz-feldspar gneiss and dark amphibolite, similar to the rocks represented in Table 3, columns 4 and 7. The only marked difference between the bulk of the volcanics of the middle-grade and high-grade zones is the slightly coarser grain size of the rocks in the latter zone. The average grain size of the matrix of the light-colored gneiss in the high-grade zone is about 0.25 to 0.5 millimeter, whereas in the middle-grade zone the average is 0.15 to 0.3 millimeter.

Among the less common rocks are sillimanite quartzite, talc-muscovite schist, and hornblende-chlorite-talc schist.

Sillimanite is present in some of the beds of quartzite (Table 3, column 10), and also in some of the light-colored gneiss, probably those containing admixed sedimentary material. The sillimanite in some of these rocks is recognizable megascopically.

A bed of talc muscovite schist (Table 3, column 9), present near the crest of Surry Mountain at the northern edge of the area, is presumably the result of retrograde metamorphism, possibly of actinolitic amphibolite.

A coarse-grained amphibolite altered to a chlorite-talc schist containing remnants of the hornblende (Table 3, column 12) has been quarried for soapstone at one locality on the west side of Sprague Brook about 1¼ miles S. 40° W. of the village of Richmond.

Thin beds of coarse anthophyllite schist occur in the Ammonoosuc volcanics in a few places in the southeastern part of the area. Although anthophyllite was identified only in the high-grade zone, its occurrence is probably controlled by the original composition of the rock.

Thickness.—The thickness of the Ammonoosuc volcanics exposed in the area decreases toward the east. About 1000 feet of this formation

TABLE 3.—ESTIMATED MODES OF THE AMMONOOSUC VOLCANICS

	Middle-grade zone										High-grade zone			
	1	2	3	4		5	6	7		8	9	10	11	12
Relic phenocrysts														
Quartz	—	—	—	7	0–30	—	5	—	—	—	—	—	15	—
Porphyroblasts														
Hornblende	—	—	—	—	—	—	6	—	—	—	—	—	—	32
Actinolite	—	—	—	—	—	13	—	—	—	—	—	—	—	—
Muscovite	—	20	—	—	—	—	—	—	—	—	23	—	—	—
Kyanite	—	—	—	—	0–2	—	—	—	—	—	—	—	—	—
Groundmass														
Quartz	78	33	50	36	11–58	22	30	2	0–10	35	2	90	51	—
Plagioclase	tr	—	29	43	35–52	62	53	39	20–50	62	12	—	—	—
Potash feldspar	—	—	15	—	tr	tr	—	—	0–tr	—	tr	—	1	—
Actinolite	—	—	—	—	0–1	—	—	—	—	—	—	—	—	—
Hornblende	—	—	—	—	0–1	—	—	58	38–70	tr	—	—	—	—
Biotite	1	—	1	4	1–8	1	4	—	tr	2	—	1	—	—
Chlorite	3	—	tr	5	1–9	1	—	tr	0–1	tr	12	—	—	22
Muscovite	15	—	4	2	0–7	—	1	—	—	—	—	1	22	—
Sericite	—	46	—	—	tr	tr	—	—	tr	—	—	—	5	—
Talc	—	—	—	—	—	—	—	—	—	—	52	—	—	38
Sillimanite	—	—	—	—	—	—	—	—	—	—	tr	8	6	—
Garnet	—	1	1	1	0–2	—	tr	—	—	—	—	—	—	—
Pyrite	1	tr	tr	—	tr	—	—	—	—	—	—	tr	—	—
Magnetite	tr	—	—	tr	tr	1	tr	1	0–1	tr	—	—	—	tr
Rutile	—	—	—	—	tr	—	—	—	—	—	—	tr	—	—
Zircon	tr	—	—	tr	tr	—	tr	—	tr	tr	tr	tr	tr	tr
Leucoxene	—	—	—	—	—	—	—	—	—	—	tr	—	—	—
Apatite	tr	—	—	tr	tr	tr	tr	tr	tr	tr	—	—	—	1
Epidote	—	tr	tr	1	0–3	—	1	—	—	tr	—	—	—	—
Carbonate	2	—	tr	—	—	—	—	—	—	—	—	—	—	—
Tourmaline	tr	—	—	—	tr	—	—	—	—	—	—	—	—	—
Sphene	—	—	—	—	—	—	tr	tr	0–1	—	—	—	—	—

Carbonate	tr	—	—	—	tr	—	—	—	—	—	—	—	—	—
Tourmaline	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Sphene	—	—	—	—	—	—	tr	tr	0–1	—	—	—	—	2
Staurolite	—	—	—	—	tr	—	—	—	—	—	—	—	—	—
Brittle mica?	—	—	—	—	—	—	—	—	—	—	—	—	—	5
Per cent of anorthite in plagioclase	—	—	9	27	19–30	26	25	30	20–40	24	26	—	—	—
Size of phenocrysts and porphyroblasts (mm)	—	0.5–0.75	—	—	2.0–15.0	1.0–5.0	1.0–2.0	—	—	—	1.0–5.0	—	5.0	5.0
Size of groundmass (mm)	0.2–0.3	0.02–0.03	0.1–0.2	0.3–0.7	0.1–0.7	0.1–0.3	0.2–0.4	1.0–2.0	0.3–2.0	0.2–0.4	0.5–1.0	0.2–0.5	0.2–0.5	0.05–0.5
Texture	Gr*	S**	Gr	Gr S	Gr S	Gr S	Gr S	S	S Gr	Gr	S	S	S	S

* Gr—Granoblastic.
** S—Schistose.

1. Muscovite-pyrite quartzite, average of two thin sections.
2. Sericite schist.
3. Light-colored quartz-feldspar gneiss.
4. Light-colored biotite-quartz-feldspar gneiss, average and range of four thin sections.
5. Light-colored actinolite-quartz-feldspar gneiss, average of two thin sections.
6. Light-colored hornblende-quartz-feldspar gneiss.
7. Amphibolite, average and range of five thin sections.
8. Pebble of light-colored biotite-quartz-feldspar gneiss in light-colored biotite-quartz-feldspar gneiss.
9. Talc-muscovite schist.
10. Sillimanite quartzite.
11. Quartzite pebble in light-colored biotite-quartz-feldspar gneiss.
12. Hornblende-chlorite-talc schist (soapstone).

is exposed in the northwestern part of the Keene quadrangle. About 750 feet of volcanics is exposed on Hyland Hill, in the north-central part of the area. In the northeastern part of the area, near Keene, only about 100 feet of this formation is present. The lower contact of the formation is an intrusive one, and beds below it are not exposed.

Partridge Formation

General statement.—In the northern half of the area, the Partridge formation crops out in an irregular belt next to the Ammonoosuc volcanics. In the southwestern and south-central parts of the area the Partridge formation is absent, but in the southeastern corner of the area it occurs sporadically at the appropriate horizon (Pl. 1). The formation is in the high-grade zone of metamorphism in the eastern fourth of the area; elsewhere it is in the middle-grade zone. Bedding is well defined in most places and is shown by numerous thin beds of quartzite or quartz-mica schist interbedded with mica schist. Bedding-plane cleavage is also very well developed.

Middle-grade zone.—In the middle-grade zone of metamorphism, the Partridge formation consists of schist and micaceous quartzite with local beds of lime-silicate rocks and volcanics.

The schists are medium-grained light-gray to medium-gray rocks composed of muscovite, biotite, quartz, and accessories. Most of these rocks contain prominent porphyroblasts of biotite, garnet, and/or staurolite. Porphyroblasts of cross-biotite are very common. Much of the schist contains sericitized plagioclase; potash feldspar is present locally. The maximum amount of feldspar found was 16 per cent. Pyrite is a very common accessory mineral and upon weathering gives the schist a characteristic greenish-yellow to brown color, in contrast to the shiny schists of the overlying Littleton formation. The grain size of these rocks averages about 0.1 to 0.25 millimeter. Biotite and garnet porphyroblasts average about 1.0 to 1.5 millimeters. In some places staurolite porphyroblasts are several centimeters long. Accessory minerals include pyrite, tourmaline, apatite, zircon, magnetite-ilmenite, and leucoxene. Estimated modes are given in Table 4, columns 2, 3, 4, 5, 6, and 9.

The micaceous quartzites are medium-gray fine-grained rocks composed of quartz, muscovite, biotite, and accessories. Many of them contain abundant porphyroblasts of garnet and biotite. An estimated mode is given in Table 4, column 1.

Light-colored biotite-quartz-feldspar gneiss (Table 4, column 8) and amphibolite, both of volcanic origin, are present in the Partridge formation. These rocks are very similar to corresponding rocks in the Ammonoosuc formation.

The lime-silicate beds, fine-grained dark-gray to gray-green rocks with coarse hornblende and garnet, represent original calcareous sediments. A lime-silicate rock from the northern slope of Cass Hill (Table 4, column 7) contains about 23 per cent plagioclase having a composition of An_{38}.

High-grade zone.—The Partridge formation is thin and discontinuous in the high-grade zone. The rocks in this zone are slightly coarser than corresponding rocks in the middle-grade zone, and locally contain some sillimanite.

Thickness.—The Partridge formation in this area thins toward the east and south and pinches out north of the Massachusetts state line. It is about 1000 feet thick in the northwestern part of the Keene quadrangle and thins to about 75 feet in the northeastern part. Schist that probably belongs to the Partridge formation is exposed on the south side of Daniels Mountain, in the west-central part of the area, but is too thin to show on the scale of the map. The variation in thickness of this formation is due, at least in part, to the unconformity at the base of the overlying formations.

Silurian Strata

Clough Quartzite

General statement.—The Clough quartzite is a thin but highly distinctive unit very useful as a key in solving the structure of the area. In the Keene-Brattleboro area this unit includes quartzite and quartz conglomerate, with occasional beds of quartz-mica schist and mica schist. Because quartzite is the most abundant rock, the name Clough quartzite has been adopted for the unit in this area.

The Clough quartzite is present in a discontinuous belt immediately above the Partridge formation or Ammonoosuc volcanics. In the

TABLE 4.—ESTIMATED MODES OF PARTRIDGE FORMATION

	1	2	3	4	5	6	7	8	9
Porphyroblasts									
Biotite	—	7	2	11	—	11	—	—	15
Garnet	2	—	—	7	2	3	4	—	5
Staurolite	—	—	—	—	—	—	—	—	tr
Hornblende	—	—	—	—	—	—	40	—	—
Groundmass									
Quartz	76	53	50	58	47	60	28	35	45
Plagioclase	10	—	11	—	—	—	23	56	16
Potash-feldspar	—	—	—	—	—	4	—	—	—
Sericite	—	—	—	—	45	—	—	—	—
Muscovite	—	30	16	13	1	18	—	2	16
Biotite	10	6	14	2	3	—	—	6	—
Chlorite	2	2	7	2	—	3	2	—	3
Garnet	—	—	—	—	—	—	—	1	—
Magnetite	tr	—	—	1	1	1	tr	tr	tr
Tourmaline	—	tr	tr	—	tr	tr	—	—	tr
Zircon	tr	tr	tr	tr	tr	—	tr	tr	tr
Apatite	tr	—	—	tr	tr	—	tr	tr	tr
Rutile	1	tr	—	—	—	—	—	—	tr
Epidote	—	—	tr	4	—	—	—	—	—
Pyrite	—	2	—	1	—	—	tr	—	—
Carbonate	—	—	—	tr	1	—	2	—	—
Sphene	—	—	—	tr	—	tr	1	—	tr
Per cent of anorthite in plagioclase	10	—	11	—	—	—	95	10	29
Grain size of porphyroblasts (mm)	0.1–0.5	0.3–0.5	2.0–4.0	0.5–1.0	0.5–1.0	1.0–2.0	3.0–5.0	—	0.5–1.5
Grain size of groundmass (mm)	0.08–0.15	0.06–0.12	0.12–0.25	0.06–0.12	0.07–0.1	0.15–0.5	0.05–0.15	0.05–0.1	0.1–0.2
Texture	Gr*	S″	S	S	Gr	S	Gr	S	S

*Gr—Granoblastic
″S—Schistose

1. Garnet-biotite quartzite.
2. Crinkled mica schist.
3. Crinkled mica-plagioclase schist.
4. Garnet-mica schist, average of 3 thin sections.
5. Garnet-sericite schist.
6. Garnet-mica schist.
7. Hornblende-anorthite granulite (lime-silicate rock).
8. Biotite-quartz-feldspar gneiss (volcanic rock).
9. Garnet-plagioclase-mica schist.

eastern fourth of the area the Clough quartzite is in the high-grade zone; elsewhere it is in the middle-grade zone.

Middle-grade and high-grade zones.—Coarse white vitreous quartzite is the most abundant rock. Garnet porphyroblasts are widespread in the quartzite, and staurolite porphyroblasts are present in a few places. Muscovite is the most abundant accessory mineral.

The quartz conglomerate, most abundant in the lower part of the formation, contains abundant pebbles of vitreous quartz in a medium-grained white quartzite matrix. Most of the pebbles have the shape of a flattened cigar. The

TABLE 5.—ESTIMATED MODES OF CLOUGH QUARTZITE

	1	2	3	4
Porphyroblasts..				
Garnet.....	—	9	1	20
Staurolite....	—	5	—	—
Groundmass....				
Quartz.....	80	72	86	12
Plagioclase...	—	—	6	3
Garnet.....	6	—	—	—
Sillimanite....	—	—	—	tr
Muscovite....	8	11	3	45
Sericite.....	tr	2	—	tr
Chlorite.....	3	—	4	9
Epidote.....	1	—	—	7
Allanite.....	—	—	—	tr
Magnetite-ilmenite....	1	1	—	1
Rutile.....	—	tr	—	tr
Leucoxene....	1	tr	—	tr
Apatite.....	tr	tr	—	1
Zircon.....	tr	tr	tr	tr
Tourmaline...	—	—	tr	2
Per cent of anorthite in plagioclase....	—	—	25	38
Size of porphyroblasts (mm).....	—	1.0–1.5	0.5–2.5	3.0–5.0
Size of groundmass (mm)...	0.3–0.5	0.2–0.3	0.2–0.4	0.5–1.0
Texture.....	Gr*	Gr	Gr	S**

* Gr—Granoblastic.
** S—Schistose.

1. Quartzite.
2. Quartz conglomerate with quartzite matrix.
3. Quartz conglomerate with quartzite matrix.
4. Muscovite-garnet schist.

ratio of the axes of the pebbles ranges up to 1:3:10. Recrystallization has been so extensive that in many cases it is hard to distinguish between the pebbles and the matrix.

The schists are medium-grained light-gray shiny rocks composed of quartz and muscovite, normally with abundant porphyroblasts of garnet as much as 6 millimeters in diameter. Accessory minerals include staurolite, tourmaline, epidote, biotite, and plagioclase.

The rocks of the Clough quartzite in the high-grade zone are very similar to those described above. Muscovite is slightly coarser, in general, and some of the more micaceous rocks contain small amounts of sillimanite.

Estimated modes are given in Table 5.

Thickness.—In the northern half of the area, the Clough quartzite is present along 40 per cent of the appropriate horizon. In the southwestern part of the area it is more nearly continuous, and in the southeastern part less continuous, than in the northern half of the area. The formation thins toward the east and southeast. Unequal original sedimentation is probably responsible for part of the variation in thickness of the formation. An erosional unconformity at the base of the overlying Littleton formation probably accounts for the absence of the Clough in places. Tectonic squeezing is certainly responsible for part of its sporadic occurrence, for pace and compass mapping demonstrates that on the limbs of many minor folds the Clough has been entirely squeezed out.

The maximum thickness of the Clough quartzite in the Keene-Brattleboro area is about 925 feet. In the northern half of the area, where measurements are most accurate, the maximum thickness is about 100 feet. The thickness of the most reliable sections is given in Table 6.

Fitch Formation

The Fitch formation was named by Billings (1935, p. 15) from a locality in the Littleton-Moosilauke area, where it overlies the Clough quartzite. The Fitch formation was recognized and mapped by Chapman (1939) in the Mascoma quadrangle, about 45 miles northeast of the Keene-Brattleboro area, where it consists mainly of lime-silicate rocks.

In the Keene-Brattleboro area, gray-green lime-silicate schist and associated purplish-brown biotite schist crop out immediately above the Clough quartzite on the 860-foot knob 2¼ miles north of the village of West Chesterfield. These rocks occur at the horizon of the Fitch formation, and they may belong to that formation. However, the Fitch formation is not distinguished on the map (Pl. 1) because of the uncertainty of the correlation.

Devonian Strata

Littleton Formation

General statement.—In the Keene-Brattleboro area the Littleton formation immediately overlies rocks of the Clough, Partridge, or Ammonoosuc formations. The Littleton is present in a belt of varying width immediately east of the Northey Hill thrust; in a narrow belt extending eastward from the village of Chesterfield; in a narrow belt extending northward from Beech Hill, near the city of Keene; in a more or less arcuate belt in the southeastern part of the area; and in a septum in the Kinsman quartz monzonite, in the central part of the area. Minerals characteristic of the high-grade zone of metamorphism are present in these rocks in the eastern part of the area and in a narrow band immediately surrounding the Kinsman quartz monzonite (Pl. 4). Elsewhere the formation displays middle-grade metamorphic minerals.

The Littleton formation consists chiefly of interbedded schist and numerous though subordinate thin quartzites. It also contains lime-silicate rocks, volcanics, and quartz conglomerate, which, where unusually extensive, are mapped separately (Pl. 1). In general, bedding is well defined throughout most of the formation.

Middle-grade zone: schist and quartzite.—These beds, originally impure shale and sandstone, are now mica schist, quartz-mica schist, and quartzite. Most of the quartz-mica schist and quartzite occur as beds a fraction of an inch to a few inches thick, interbedded with mica schist (Pl. 6, fig. 2).

The schist and interbedded quartzites are light-gray to dark-gray fine-grained to medium-grained rocks that have a distinct silvery luster. The essential minerals are quartz, muscovite, and biotite, in different proportions; plagioclase is locally present. The average grain size of the groundmass is about 0.05 to 0.25 millimeter.

Table 6.—Thickness of the Clough Quartzite in the Keene-Brattleboro Area

Location	Thickness (Feet)	Remarks
1⅛ miles northeast of Park Hill	95	Good control top and bottom
1⅛ miles southwest of Corner School	90	Good control top and bottom
1 mile northeast of Chesterfield village	40	Good control top and bottom, some folding
Head of California Brook	42	Good control top and bottom
West side of Beech Hill, Keene	60	Good control at top
¼ mile northeast of Richmond village	35	Maximum thickness exposed; exposure is on steep slope
East side of 1320 hill, 300 feet east of Pudding Hill Road	55	No folding where section measured; good control
South of Lily Pond, near Vermont-Mass. state line	190	Some folding
Bear Mountain, due east of Porter Hill	190	Good control top and bottom; thickened by pegmatite
¼ mile due west of The Oxbow	125	Minimum; good control at base
¼ mile west of Indian Pond	535	Fair control, some folding near top, may be repeated by faulting
East side of Indian Pond	925	Fair control; may be repeated by faulting

Porphyroblasts of biotite, garnet, and/or staurolite are present in most of these rocks. The biotite and garnet reach a diameter of a few millimeters, whereas staurolite has a maximum length of 30 to 40 millimeters. In some localities, the porphyroblasts have been altered to chlorite or sericite during retrograde metamorphism. Accessory minerals include tourmaline, zircon, apatite, sphene, magnetite-ilmenite, and leucoxene. Estimated modes are given in Table 7, columns 3 to 6. Tourmaline is especially abundant in the schist along the western side of the body of Kinsman quartz monzonite, where some of the rock is so highly tourmalinized that the name tourmaline schist is appropriate.

Middle-grade zone: quartz conglomerate and quartzite.—The quartzite and quartz conglomer-

TABLE 7.—ESTIMATED MODES OF THE LITTLETON FORMATION

	Middle-grade zone																High-grade zone		
	Quartzite		Mica schist				Lime-silicate and associated rocks								Volcanics		Schist		
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19
Porphyroblasts																			
Biotite	—	—	4	5	—	9	—	—	—	—	—	—	18	—	—	—	—	—	—
Chlorite	—	—	—	4	—	3	—	—	—	—	—	—	—	—	—	—	—	—	—
Garnet	—	2	3	3	5	4	—	3	—	—	—	—	—	—	—	—	—	—	—
Staurolite	—	—	—	tr	20	3	—	—	—	—	—	—	—	—	—	—	—	—	—
Hornblende	—	—	—	—	—	—	—	—	—	—	—	—	—	2	—	—	—	—	—
Sillimanite	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	12	—
Leucoxene	—	—	3	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Groundmass																			
Quartz	78	84	53	50	25	69	70	72	58	30	5	16	5	30	40	42	24	35	37
Plagioclase	14	8	—	10	—			18	20	—	35	—	10	46	44	25	7	16	
Potash-feldspar	—	—	—	—	—	—	—	—	tr	tr	—	—	—	—	—	25	tr	—	—
Muscovite	3	2	21	27	30	12	6	tr	1	2	—	—	—	—	—	1	30	2	35
Biotite	1	3	15	—	20	—	20	6	1	tr	—	—	—	21	3	5	30	32	25
Chlorite	3	1	tr	—	—	—	3	tr	tr	—	—	3	2	tr	12	—	tr	—	—
Garnet	1	—	—	—	—	—	1	—	9	—	—	—	—	1	tr	tr	8	2	3
Actinolite	—	—	—	—	—	—	—	—	4	58	60	58	62	—	—	—	—	—	—
Diopside	—	—	—	—	—	—	—	—	6	—	—	—	—	—	—	—	—	—	—
Staurolite	—	tr	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Sillimanite	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	tr	tr	tr
Zircon	tr	tr	—	—	—	tr	tr	tr	tr	tr	—	tr	tr	tr	—	tr	tr	tr	tr
Apatite	tr	tr	—	—	tr	tr	tr	tr	—	tr	tr	—	tr	tr	—	—	tr	tr	tr
Sphene	—	—	tr	—	—	tr	—	—	tr	—	tr	1	—	—	—	—	—	tr	—
Carbonate	—	—	—	—	—	—	tr	—	—	—	—	—	—	tr	—	—	—	—	—
Magnetite-ilmenite	tr	tr	1	1	tr	tr	—	tr	—	—	—	—	3	—	1	tr	—	1	—
Tourmaline	tr	tr	tr	—	—	tr	tr	tr	—	—	—	tr	—	tr	—	—	—	—	—
Zoisite	—	—	—	—	—	—	tr	—	1	tr	tr	22	tr	—	—	2	—	—	tr

Tourmaline	tr	tr	tr	—	—	—	tr	—	1	tr	tr	22	tr	—	—	2	—	—	tr
Zoisite	—	—	—	—	—	—	tr	—											
Pyrite	—	—	—	—	—	—	—	—	tr	2	tr	—	—	—	—	—	—	—	—
Alteration	—	—	—	—	—	—	—	—	—	8	—	—	—	—	—	—	—	—	—
Per cent of anorthite in plagioclase	6	22	—	13	—	18	20	20	83	—	25	—	a″	26	17	28	32	26	30
Grain size of porphyroblasts in mm	—	2.0	1.0–2.0	0.5–2.0	2.0–10.0	1.0–6.0	—	0.5	—	—	—	—	1.0	1.0	—	—	—	5.0–15.0	—
Grain size of groundmass in mm	0.1–0.25	0.2–1.0	0.06–0.15	0.07–0.15	0.25	0.15–0.5	0.05–0.5	0.15–0.3	0.15–0.3	0.1–1.0	0.05–0.3	0.1–0.5	0.05–0.25	0.1–0.2	0.03–0.06	0.05–0.1	0.5–2.0	0.2–0.5	0.3–0.6
Texture	Gr*	Gr	S**	S	S	S	S	Gr	Gr	Gr	S	Gr	Gr	S	Gr	Gr	S	S	S

1. Banded feldspathic quartzite.
2. Quartz conglomerate with quartzite matrix.
3. Garnet-mica schist.
4. Plagioclase-garnet-mica schist.
5. Staurolite-garnet-mica schist.
6. Staurolite-garnet-mica schist, average of 2 thin sections.
7. Purplish-brown biotite schist.
8. Feldspathic quartzite.
9. Bytownite-garnet-diopside granulite.
10. Actinolite-quartz granulite.
11. Actinolite-plagioclase schist.
12. Actinolite-zoisite granulite.
13. Actinolite-biotite gneiss.
14. Biotite-hornblende gneiss.
15. Laminated biotite granulite (originally tuff).
16. Biotite-quartz-feldspar gneiss.
17. Garnet-sillimanite schist.
18. Sillimanite schist.
19. Garnet-sillimanite schist.

ate of the Littleton formation are discontinuous and have a maximum thickness of only a few tens of feet. Most of them occur near the base of the formation. These rocks are white to light-gray medium-grained quartzite and quartz-conglomerate with small amounts of mica and other accessories. The pebbles in the conglomerate are ellipsoidal and range up to 8 or 10 inches in length.

Microscopic examination shows a granoblastic texture and an average grain size of 0.2 to 1.0 millimeter. Essential minerals are quartz and mica; accessory minerals include garnet, tourmaline, apatite, magnetite, and zircon. Plagioclase, locally altered to sericite, is present in some of the quartzite. Garnet porphyroblasts with a maximum diameter of about 2 millimeters occur in both quartzite and quartz conglomerate. Estimated modes are given in Table 7, columns 1 and 2.

Volcanics and lime-silicate rocks are locally associated with the quartz conglomerate.

Middle-grade zone: lime-silicate and associated rocks.—The lime-silicate beds of the Littleton formation include actinolite schist, actinolite-quartz granulite, bytownite-garnet-diopside granulite, and similar rocks. These beds were originally arenaceous dolomites and dolomitic shales. Associated with these lime-silicate rocks are feldspathic quartzite and purplish-brown biotite schist.

The actinolite schists are light-green to gray-green weakly to strongly foliated rocks. Actinolite needles lie in the plane of foliation and locally show alignment. Microscopic examination shows that these rocks have an average grain diameter of about 0.1 to 0.3 millimeter. Principal minerals include actinolite, quartz, plagioclase, and zoisite; accessories include sphene, tourmaline, zircon, and apatite.

Individual beds of granulite are in some cases only a few millimeters thick. Bedding is shown by changes in color through white, gray, pink, green, and brown, due to differences in mineral composition. Some beds are rich in actinolite, others are rich in diopside, calcic plagioclase, biotite, muscovite, quartz, or garnet. These rocks have a granoblastic texture and an average grain size of 0.15 to 0.5 millimeter. Zoisite, sphene, zircon, and pyrite are included in the accessory minerals.

The feldspathic quartzites are light-gray well-bedded rocks with a sugary texture. They probably represent original arkosic sandstones, which now contain quartz, oligoclase, biotite, and muscovite, with the usual accessories.

Purplish-brown biotite schists, with which the other rocks are interbedded, make up the greater part of the lime-silicate members. These rocks have a schistose to granoblastic texture, with grains averaging 0.05 to 0.5 millimeter in diameter. Quartz, biotite, muscovite, and plagioclase are the essential minerals. These schists are strikingly different in appearance from the usual silvery mica schist of the Littleton formation because of their purplish-brown color and their tendency toward a granoblastic texture.

The biotite in the lime-silicate rocks is high in magnesium and low in iron compared to the biotite of the argillaceous rocks.

Estimated modes of the lime-silicate and associated rocks are given in Table 7, columns 7 to 14.

Middle-grade zone: volcanics.—Light-colored biotite-quartz-feldspar gneiss and amphibolite occur at numerous horizons in the Littleton formation. Amphibolite, originally basalt or andesite, is somewhat more abundant than the light-colored volcanics. Some of the amphibolites were originally bedded tuffs; other more massive varieties probably represent flows; and still others, because of the contact relations, are interpreted as sills and dikes. The biotite-quartz-feldspar gneiss, inasmuch as it is bedded, probably represents metamorphosed tuffs, of intermediate composition. In hand specimen, the volcanics of the Littleton formation are very similar to those of the Ammonoosuc formation.

Estimated modes are given in Table 7, columns 15 and 16.

High-grade zone: schist and quartzite.—The schist and interbedded quartzites of the high-grade zone are medium-grained to coarse-grained rocks, coarser than corresponding rocks in the middle-grade zone. Many of these rocks contain abundant porphyroblasts of garnet and sillimanite. The sillimanite, which attains a maximum length of 50 millimeters, was locally replaced by sericite during retrograde metamorphism. Microscopic examination shows that these rocks have an average grain size of 0.2 to

2.0 millimeters. Plagioclase (An_{28} to An_{32}) is conspicuous. Estimated modes are given in Table 7, columns 17, 18, and 19.

High-grade zone: quartz conglomerate and quartzite. —The quartz conglomerate and quartzite of the high-grade zone are very similar to corresponding rocks in the middle-grade zone. Further recrystallization has resulted in a slightly coarser grain size, and the more impure quartzites contain traces of sillimanite.

High-grade zone: lime-silicate rocks.—Although the appropriate horizons are exposed, no lime-silicate beds were observed in the high-grade zone.

High-grade zone: volcanics.—Most of the volcanics of the Littleton formation in the high-grade zone are coarse amphibolites. Well-bedded light - colored biotite - quartz - feldspar gneiss is associated with the amphibolite in some places. These rocks are similar to corresponding types in the middle-grade zone.

Thickness and stratigraphy of lithologic units. —Approximately 5000 feet of the Littleton formation is exposed in the Keene-Brattleboro area. This is the youngest formation in the area; its top is not exposed.

Thin beds of quartzite and quartz conglomerate are exposed in several places near the base of the Littleton. As these rocks occur in discontinuous lenses, it is not known whether they are all at the same horizon. The quartzite shown on Hyland Hill, in the north-central part of the area, (Pl. 1) has a maximum thickness of about 20 feet and is about 450 feet above the base of the Littleton. A quartz conglomerate about 10 feet thick is associated with the lens of volcanics just east of Fox Hill near the village of South Vernon (Pl. 1) and is about 390 feet above the base of the Littleton. Quartzite and quartz conglomerate at a much higher horizon are exposed on Whipple Hill, in the southeastern part of the area. This bed, a few tens of feet thick, is between 2000 and 2500 feet above the base of the Littleton.

Volcanics are present at numerous horizons throughout the Littleton. Three such zones are mapped in the southeastern part of the town of Putney. The lower zone is about 200 feet thick, the two upper zones are each about 100 feet thick; they occur about 700 feet, 1100 feet, and 1300 feet, respectively, above the base of the formation. The lens of felsitic volcanics in the southeastern part of the town of Hinsdale, near the village of Doolittle, has a maximum thickness of 1400 feet and is about 3000 feet above the base of the formation.

Lime-silicate rocks are also present at several horizons. The zone of lime-silicate rocks in the southwestern part of the township of Chesterfield has a maximum thickness of about 1200 feet and is about 2500 feet stratigraphically above the base of the formation.

Age and Correlation of Strata

No fossils have been found in the Keene-Brattleboro area. Correlations are based on the tracing of map units into areas which have been mapped, on lithologic similarities with rocks of other areas, and on similarities of the stratigraphic column with those of other areas.

The Waits River formation was identified by its characteristic beds of blue-gray marble and by tracing it by reconnaissance mapping northeastward to the Bellows Falls area where it has been identified by Kruger (1946). These calcareous beds, in the past, have been assigned to the Ordovician on the basis of supposed graptolites reported by Richardson (1919); more recently these "graptolites" have been interpreted as linear streaking of mica flakes (Foyles, 1931). Currier and Jahns (1941, p. 1492) state that near Northfield, Vermont (Fig. 1, 16 miles northeast of Lat. 44°, Long. 73°), the base of the Waits River formation is approximately 2250 feet stratigraphically above a thin crinoidal limestone; and, as crinoids did not appear until the latest Cambrian, the Waits River is Ordovician or younger. According to Josiah Bridge and P. E. Raymond, the size and relative abundance of the crinoids in the limestone suggest that they are not older than middle Ordovician (Currier and Jahns, 1941, p. 1500–1501). The Waits River formation, therefore, is middle Ordovician or younger.

Structural evidence in the Keene-Brattleboro area indicates that the top of the Waits River formation is toward the east. There appears to be no major stratigraphic or structural break between the Waits River formation and the Orfordville formation, and none within the Orfordville formation. Beds mapped as the lower mem-

ber of the Orfordville are lithologically identical with beds that Kruger (1946) mapped as the "Lower Schist Member" of the Orfordville, and with part of the Gile Mountain formation as mapped by Doll (1944). Thus the Orfordville formation appears to overlie the Waits River formation, and it, also, would be middle Ordovician or younger.

The Orfordville formation in its type area, the Mt. Cube quadrangle, underlies the Albee and Ammonoosuc formations (Hadley, 1942) and was assigned tentatively to the Ordovician. Although the Orfordville formation has not been traced by continuous detailed mapping from the type locality to the Keene-Brattleboro area, the similarity in the lithology and in the structural position of the unit in the two areas indicates that the correlation is valid.

Doll (1943a) reports fossils from the Waits River formation at Westmore, Vermont (Fig. 1, F?, at Lat. 44°45′, Long. 72°), that are at least as young as middle Silurian and are very possibly of early Devonian age. Five well-known paleontologists identified some of the forms as echinoids, one qualified his identification, and two equally well known paleontologists doubted the organic origin of the forms. Doll (1943b) also reports a single determinable brachiopod from the Gile Mountain formation on Whitcomb Hill, South Strafford, Vermont (Fig. 1, F? at Lat. 43°50′, Long. 72°20′). He identified the form as probably *Spirifer murchisoni* and believed it indicated an early Devonian (Oriskany) age for the formation. White (1946, p. 133) has seen this "brachiopod" and thinks it is a piece of fractured vein quartz. Ages assigned on the basis of these "fossils" can hardly be considered as proven.

The Waits River and Orfordville formations are separated from the remainder of the rock units, in this area, by the Northey Hill thrust.

The ages of the formations east of the Northey Hill thrust are based on fossils found by Billings and Cleaves (1934) in the Littleton-Moosilauke area. Accurate correlation between the rocks of that area and those of the Keene-Brattleboro area is possible because the intervening quadrangles have been mapped in detail. The Clough quartzite can be traced continuously from the Moosilauke quadrangle to the Keene quadrangle with only one small gap, in the Bellows Falls quadrangle, where plunging structures interrupt the continuity. Map units in the Keene-Brattleboro area are very similar to corresponding units in the Littleton-Moosilauke area.

In the Littleton-Moosilauke area (Fig. 1), the Fitch formation, which overlies the Clough, contains fossils of middle Silurian age (Billings and Cleaves, 1934). The Clough formation, because it is closely associated with the Fitch, is considered to be middle or lower Silurian. The Albee, the Ammonoosuc, and the Partridge, which underlie the Clough, are considered to be Ordovician (?). The Littleton formation, which overlies the Fitch formation, contains fossils of early Devonian (Oriskany) age (Billings and Cleaves, 1934).

The Bernardston formation of Massachusetts, named for its occurrence in the town of Bernardston (Dana, 1873), is continuous with the Littleton formation of New Hampshire. A well-known fossil locality in the Bernardston formation, 0.5 mile northwest of the village of Bernardston (Fig. 1, Bernardston quadrangle), is about 4 miles south of the Keene-Brattleboro area.

Fossils collected from the Bernardston locality have been variously dated as middle Silurian to upper Devonian. Emerson (1898, p. 254) summarizes the age assignments of the Bernardston formation up to the date of his paper. E. Hitchcock (1851, p. 299), on the basis of crinoids identified by James Hall, assigned the whole of the Bernardston formation to the Devonian. Hall thought the crinoids, which came from a bed of limestone in the Bernardston formation, where the same age as the Onandaga limestone. Whitfield (1883, p. 368), on the basis of the fossils that he examined, concluded that the limestone may be of middle Silurian age and that a thin fossiliferous quartzite immediately overlying the limestone is most probably Devonian. He thought the fossils from the quartzite resembled those from the New York Chemung group (upper Devonian). John M. Clarke (*in* Emerson, 1898, p. 259), after examining fossils from this locality, found that the forms in the limestone and in the quartzite were the same and concluded that the fauna is late Devonian in age.

Schuchert and Longwell (1932, p. 321, 322) make the following positive statements concerning the age of the Bernardston formation, "... the fossiliferous Devonian of the Connecticut Valley at Bernardston, Massachusetts, since the latter is of late Lower Devonian age ..."; and "Finally at Bernardston, in the Connecticut Valley, which is east of the Green Mountain axis, unmistakable late Lower Devonian fossils have also been found . . .".

Robert Balk (1941, p. 2009) states, "The fossils available in various collections have been reexamined recently by Dr. G. A. Cooper of the U. S. National Museum, and one single individual, *Spirifer divaricatus*, is determinable specifically and fixes the age of the marble as middle Devonian (Onondaga)". As there is disagreement on the age of the Bernardston, it will be necessary to do more paleontological work to evaluate the significance of this one fossil.

Geologic mapping shows that the Littleton formation is continuous with the Bernardston formation. Correlation between the Littleton formation, of lower Devonian Oriskany age (Billings and Cleaves, 1934), and the Bernardston formation, on the basis of fossils, is good insofar as the Bernardston formation is certainly Devonian. Accepting the statements of Schuchert and Longwell on the age of the Bernardston, fossil evidence shows that both formations are lower Devonian (Oriskany).

Intrusive Igneous Rocks

General Statement

Intrusive igneous rocks cover about 45 per cent of the Keene-Brattleboro area. Foliated quartz monzonite, granodiorite, and quartz diorite are by far the most abundant of these plutonic rocks. Following the classification of Billings (1937), they belong to the Oliverian magma series and the New Hampshire magma series. The rocks of the Oliverian magma series, either pretectonic or early tectonic in age, occupy the cores of several domes. The New Hampshire magma series, intruded during later stages of the orogeny, occurs as sheets, stocks, sills, and dikes. A group of amphibolite sills and dikes are probably early members of the New Hampshire magma series. One highly weathered post-metamorphism mafic dike, which may belong to the White Mountain magma series, was found in the area. Pegmatites are abundant and are associated with both the Oliverian magma series and the New Hampshire magma series. The largest pegmatites occur in the northeastern part of the area, where some of them are quarried.

Oliverian Magma Series

General statement.—Billings (1935, p. 26) introduced the term Oliverian magma series to designate a comagmatic group of intrusive rocks in western New Hampshire. The rocks of this magma series, intrusive into the surrounding metamorphic rocks, form the cores of elongate domical uplifts.

The Oliverian magma series is represented by three plutons in the Keene-Brattleboro area. The Westmoreland-Swanzey pluton, in the northern and eastern parts of the area, is the largest and underlies most of the towns of Westmoreland, Surry, Keene, Swanzey, and Richmond, and part of Chesterfield. The Vernon pluton is present in the southwestern part of the area in the towns of Vernon and Hinsdale. The northern part of the Warwick pluton, so named because most of it is in the town of Warwick, Massachusetts, is present in the south-central part of the area.

Foliation is well developed around the edges of each of these plutons but decreases somewhat in perfection toward their centers. In general, the foliation in each pluton is parallel, in plan, to the contact of the gneiss with the overlying Ammonoosuc volcanics.

Gneiss of the Westmoreland-Swanzey pluton.—The Westmoreland-Swanzey pluton is nearly divided by tongues of metamorphic rock mmediately west of the city of Keene; the two lobes will be referred to as the northwestern lobe and the eastern lobe. The northwestern lobe is about 10 miles wide and 7 miles long; the eastern lobe is about 7 miles wide and 16 miles long. Two prongs of the northwestern lobe extend northward into the Bellows Falls quadrangle.

Most of the gneiss in the northwestern lobe is granodiorite; the northwestern part, however, is quartz monzonite, and local areas of granite, not separately mapped, are present. The granodiorite of the southeastern part of this lobe contains hornblende in addition to the more abundant biotite.

Most of the gneiss of the eastern lobe of the pluton is quartz diorite; local areas are richer in potash feldspar. Much of this gneiss contains both biotite and hornblende.

The contacts between these varieties of gneiss are gradational, and presumably the varieties of gneiss are contemporaneous.

The different varieties of gneiss are very similar in hand specimen. They are light-gray to salmon-colored medium-grained granular biotite gneisses. Potash feldspar gives the rock its salmon color, hence the granite and quartz monzonite tend to be pinker than the granodiorite and quartz diorite. The rock is locally subporphyritic, with granulated and flattened eyes of milky quartz as much as 15 millimeters across, and occasional pink microcline phenocrysts. The hornblende, where present, also occurs as phenocrysts. In the groundmass, plagioclase, potash feldspar, quartz, biotite, epidote, garnet, and magnetite can be recognized megascopically. Light-green epidote is commonly associated with the biotite. Magnetite in aggregates up to 20 millimeters in diameter or as smaller single grains is characteristic of the Oliverian magma series, and can be found in almost every outcrop. Occasional small pink garnets, 0.1 to 0.5 millimeter in diameter, are present in much of the gneiss. Clusters of biotite up to 25 millimeters across are also characteristic of the gneiss. These clusters are circular or slightly elongate aggregates of biotite flakes lying along foliation planes (Pl. 6, fig. 4). The long axes of the elongate biotite clusters and the quartz eyes locally show parallel orientation, thus producing a marked linear feature.

Microscopic examination of the gneiss shows a holocrystalline subporphyritic rock with hypidiomorphic to granoblastic texture. The quartz eyes show granulation and undulatory extinction; microcline phenocrysts are ungranulated. Ungranulated phenocrysts of hornblende, 30 millimeters in maximum length, poikilitically enclose minerals of the groundmass. The grain size of the groundmass averages 0.5 to 1.0 millimeter. The essential minerals are plagioclase (An_{12} to An_{48} in different specimens), potash feldspars (microcline, orthoclase, and some microperthite), quartz, and biotite. Accessory minerals include magnetite, garnet, epidote, muscovite, apatite, zircon, sphene, and rutile. Some chlorite and sericite are present as alteration products. Characteristic modes are given in Table 8.

The average gneiss of the pluton contains 4 to 6 per cent biotite and chlorite. Locally, the gneiss is almost white; such rocks are similar to the average gneiss of the body except that they contain only a trace of biotite, but more than the average amount of magnetite.

Small pegmatites are associated with the gneiss and occur either in shear zones or in dikelike and sill-like masses. Many shear zones, along which foliation shows slight offset, are filled with coarse-grained pegmatitic pink microcline, quartz, biotite, and magnetite (Pl. 9, fig. 1). The pegmatitic filling of the shear zones grades into the country rock. The minerals listed above also occur in the dikelike and sill-like pegmatites.

Thin schlieren of fine-grained biotite-rich gneiss (Pl. 8, fig. 2), probably reworked inclusions of amphibolite of the Ammonoosuc volcanics, are most abundant near the border of the intrusive. These schlieren have a granular texture, an average grain size of 0.1 to 0.2 millimeter, and contain 10 to 15 per cent biotite—more than the average gneiss contains. They contain few quartz eyes, and the biotite does not occur in clusters.

In addition to the main body of the pluton, there are many sills of this gneiss in the overlying Ammonoosuc volcanics. A few are mapped, but most of them are too small to be shown on the scale employed. In many places near the contact with the Ammonoosuc, the gneiss contains abundant inclusions of the volcanics (Pl. 5, figs. 3, 4). In many cases, it is not clear whether certain outcrops should be mapped as gneiss, full of inclusions, or as Ammonoosuc volcanics that contain abundant sills.

Gneiss of the Vernon pluton.—The Vernon pluton, an elongate oval body about 8 miles long and 2 miles wide, occupies the lowland area between the southern end of Wantastiquet Mountain and the Vermont-Massachusetts state line. The Connecticut River flows southeasterly across the central part of the body.

The gneiss of this pluton, most of which is quartz diorite, is a light-gray foliated subporphyritic rock, very similar to the gneiss of the other two plutons of Oliverian magma series in this area. As shown in the estimated modes (Table 9), the gneiss contains plagioclase (An_{29}

TABLE 8.—ESTIMATED MODES OF ROCKS OF OLIVERIAN MAGMA SERIES, WESTMORELAND-SWANZEY PLUTON

	1		2		3	4	5		6
Phenocrysts									
Quartz	4	0–20	—	0–5	—	—	tr	0–2	—
Hornblende	tr	0–2	—	—	—	—	5	4–7	—
Groundmass									
Plagioclase	44	33–55	31	27–35	21	18	53	52–54	39
Potash-feldspar	11	5–15	28	22–35	40	45	1	0–3	22
Quartz	33	30–42	33	30–40	35	30	32	30–35	36
Biotite	2	tr–6	5	3–8	tr	3	5	1–8	2
Chlorite	1	0–4	0.5	0–2	tr	1	1	0–2	tr
Muscovite	2	0–7	2	1–3	4	3	—	—	tr
Epidote	2	0–10	tr	tr	tr	tr	1	1	—
Magnetite	tr	tr	0.5	tr–1	tr	1	tr	tr	1
Garnet	tr	tr	tr	tr–1	tr	—	tr	0–1	tr
Zircon	tr	tr	tr	tr	tr	tr	tr	tr	tr
Apatite	tr	tr	tr	tr	tr	tr	tr	tr	—
Sphene	—	tr	—	—	—	—	—	tr	—
Pyrite	—	tr	—	tr	tr	—	—	—	—
Hematite	—	0–2	—	—	—	—	—	—	—
Allanite	—	—	—	—	—	—	tr	tr	—
Per cent of anorthite in plagioclase	26	17–41	27	24–32	12	20	45	41–48	20
Grain size of groundmass (mm)	0.2–1.0	0.25–1.5	0.5–1.0	0.25–1.5	0.3–1.0	0.2	1.0–1.5	0.7–2.0	0.75–1.0
Grain size of phenocrysts (mm)	4.0–6.0	4.0–6.0	6.0–12.0	6.0–15.0	—	—	3.0–15.0	3.0–15.0	—
Texture	Gr* Subp**	Gr Subp	Gr	Gr Subp	Gr	Gr	Gr Subp	Gr Subp	Gr

* Gr—Granoblastic.
** Subp—Subporphyritic.

1. Grandiorite, average and range of 9 sections.
2. Quartz monzonite, average and range of 4 sections.
3. Quartz monzonite-granite, average of 2 sections.
4. Granite, 1 section.
5. Quartz diorite, average and range of 3 sections.
6. Quartz monzonite, 1 section.

to An_{44}), quartz, and biotite as essential minerals. Epidote, muscovite, potash feldspar, hornblende, garnet, and magnetite are included in the list of accessory minerals; sericite, chlorite, and calcite are present as alteration products. Granulated and flattened quartz eyes are present; individual grains of these aggregates show sutured and mortar textures. The quartz eyes, and strings of biotite flakes, locally produce a prominent lineation.

In general, the gneiss of the Vernon pluton shows slightly more granulation and alteration than does the gneiss of the other two plutons of the Oliverian magma series.

Inclusions of amphibolite and fine-grained biotite gneiss of the Ammonoosuc volcanics, some of which are discordant, are abundant in places; and numerous sills of quartz diorite occur in the overlying Ammonoosuc volcanics. Hornblende is locally concentrated around some of the amphibolite inclusions.

Gneiss of the Warwick pluton.—The Warwick pluton, in the Keene-Brattleboro area, occupies about 9 square miles of lowland area immediately south of the village of Winchester. Most of this pluton is in the adjoining Northfield and Mt. Grace quadrangles, Massachusetts, where it has been mapped by Robert Balk and Jarvis B. Hadley.

The Warwick pluton, in this area, consists al-

TABLE 9.—ESTIMATED MODES OF THE OLIVERIAN MAGMA SERIES

	Vernon pluton					Warwick pluton	
	1	2	3	4	5	6	7
Phenocrysts							
Quartz	—	2	5	8	—	—	—
Plagioclase	—	2	—	2	—	—	—
Hornblende	—	—	—	5	—	—	—
Groundmass							
Plagioclase	54	52	58	54	53	50	49
Potash-feldspar	tr	tr	—	—	7	11	13
Quartz	32	36	31	30	36	36	33
Biotite	5	3	—	—	3	3	4
Chlorite	2	1	2	tr	—	—	—
Muscovite	3	2	4	—	tr	tr	tr
Epidote	4	1	—	—	1	1	1
Magnetite	—	tr	tr	tr	tr	—	tr
Garnet	—	tr	tr	—	—	tr	tr
Zircon	tr	tr	tr	tr	tr	tr	tr
Apatite	tr	tr	tr	tr	tr	tr	tr
Sphene	—	—	—	—	tr	—	—
Pyrite	—	—	tr	—	—	—	—
Hematite	—	—	tr	—	—	—	—
Leucoxene	—	—	tr	—	—	—	—
Tourmaline	—	—	tr	—	—	—	—
Carbonate	tr	1	—	—	—	—	—
Per cent of anorthite in plagioclase	41	37	12	44	41	29	28
Size of groundmass (mm)	0.25–1.0	0.1–0.25	0.05–0.2	0.15–0.3	0.05–0.5	0.5–1.0	0.7–1.2
Size of phenocrysts (mm)	—	2.0–3.0	2.0–4.0	1.0–4.0	—	—	—
Texture	Gr*	Gr Subp″	Gr Subp M**	Gr Por′	Gr M	Gr	Gr

* Gr—Granoblastic
″ Subp—Subporphyritic
** M—Mortar
′ Por—Porphyritic

Vernon pluton
1. Quartz diorite
2. Quartz diorite
3. Quartz diorite
4. Hornblende quartz diorite
5. Granodiorite

Warwick pluton
6. Granodiorite
7. Granodiorite

most entirely of well-foliated medium-grained gray granodiorite gneiss very similar to that in the Westmoreland-Swanzey pluton. The gneiss is composed of plagioclase, quartz, potash feldspar, and biotite; magnetite and garnet are among the accessory minerals. Much of the gneiss in the eastern part of the body contains as much as 5 per cent hornblende. Parts of the gneiss are subporphyritic, and elongate quartz eyes and biotite clusters are locally aligned to produce a prominent mineral lineation. Estimated modes of the granodiorite are given in Table 9.

Shear zones containing a pegmatitic filling are locally present and are similar to those in the Westmoreland-Swanzey pluton.

Numerous inclusions of fine-grained to coarse-grained amphibolite, the largest of which is about 1000 feet across, are present in the granodiorite (Pl. 1). Small fine-grained dark-gray schlieren that contain 15 to 20 per cent biotite are also present in the granodiorite. These schlieren probably represent reworked inclusions of amphibolite.

Numerous sills of granodiorite occur in the overlying Ammonoosuc volcanics.

New Hampshire Magma Series

General statement.—The New Hampshire magma series is represented in the Keene-Brattleboro area by thin sills and dikes of porphyritic amphibolite, by numerous small stocks and sills of granodiorite and quartz diorite, by the Kinsman quartz monzonite, by the Spaulding quartz diorite, and by the Concord granite. As most of the bodies are not in contact with other members of this series, their relative ages are based on the amount of foliation, shearing, and granulation that they show.

Sills and dikes of amphibolite are most numerous in the southeastern part of the area. Small stocks and sills of granodiorite and quartz diorite occur in many scattered localities. A pluton of Kinsman quartz monzonite is present in the central part of the area. The Spaulding quartz diorite occurs in a small irregularly shaped body along the east-central edge of the area. The Concord granite occurs in an elongate body in the northeastern corner of the area and as small stringers in the surrounding schist.

Amphibolite sills and dikes.—Sills and dikes of amphibolite, originally basaltic or andesitic intrude rocks of the Oliverian magma series in the southern part of the eastern lobe of the Westmoreland-Swanzey pluton. A special symbol is used for those areas where amphibolite sills and dikes are most numerous (Pl. 1).

The amphibolite is a well-foliated medium-grained dark rock containing, in most places, granulated and drawn-out phenocrysts of white plagioclase (Pl. 7, fig. 1). Most of the amphibolite occurs in sills; small apophyses extend from some sills into the country rock. Though a few of the sills are several tens of feet thick, the thinner ones are better exposed. Many of these bodies show chilled border zones of fine-grained nonporphyritic amphibolite.

The foliation in both sills and dikes is parallel to the foliation in the enclosing rock. The foliation in the sills is parallel to the contacts of the sills. The foliation in the dikes cuts across the contacts of the dikes. A small dike showing such relations is exposed on the east slope of Mount Caesar, in Swanzey Township, at an elevation of about 790 feet. The foliation in the amphibolite is a secondary foliation induced by regional deformation.

A typical sill of porphyritic amphibolite, exposed in he steep cliff on the northwest side of Swanzey Lake in the gneiss of the Oliverian magma series, is about 5 feet thick. The contacts of the sill parallel the foliation of the gneiss. The lower foot and the upper 10 inches of the sill are nonporphyritic chilled borders. The porphyritic center of the sill contains about 25 per cent plagioclase phenocrysts 5 to 10 millimeters long. The phenocrysts are granulated and flattened in the plane of the foliation, which is parallel to that of the enclosing gneiss. Microscopic study shows that recrystallization of the individual grains of the granulated phenocrysts has erased textural evidence of the granulation. Essential minerals of the amphibolite are labradorite and hornblende; sphene is an abundant accessory (Table 10, column 2). The small amount of pyroxene present replaces hornblende. Quartz is intergrown with the pyroxene to produce a vermicular texture.

These amphibolites are younger than the rocks of the Oliverian magma series, but, as they are foliated and contain granulated phenocrysts, they were intruded before the end of the orogeny. Chilled contacts between the amphibolite and the gneiss of the Oliverian magma series indicate that the gneiss was cool before the amphibolites were intruded and that considerable time elapsed between the intrusion of the two types of rock.

Granodiorite and quartz diorite.—Numerous small bodies of binary granodiorite and quartz diorite are present in the Keene-Brattleboro area. Some of these bodies are small stocks with discordant contacts in some places and concordant contacts in other places; some are lenses that are concordant in general but crosscutting in detail, and others are thin sills. In most places foliation is weak or absent, but locally along the borders of these bodies it is well developed. In

TABLE 10.—ESTIMATED MODES OF THE NEW HAMPSHIRE MAGMA SERIES

	1	2	3	4	5	6	7	8	9
Phenocrysts									
Quartz	—	—	—	10	—	—	1	—	—
Plagioclase	5	25	—	—	—	tr	4	—	1
Groundmass									
Quartz	tr	tr	33	13	26	18	30	32	38
Plagioclase	40	23	45	63	67	70	59	59	49
Potash-feldspar	—	—	17	6	—	1	tr	—	—
Hornblende	53	48	—	—	—	6	—	—	—
Pyroxene	tr	2	—	—	—	—	—	—	—
Biotite	tr	—	—	5	4	2	—	—	tr
Chlorite	—	—	2	1	1	3	—	—	2
Muscovite	—	—	1	1	1	tr	6	9	8
Sericite	—	—	tr	—	—	tr	—	—	—
Epidote	—	tr	2	1	tr	tr	—	—	2
Garnet	—	—	—	—	tr	—	—	tr	—
Sphene	2	2	tr	tr	—	tr	—	—	—
Apatite	tr	tr	tr	—	tr	tr	tr	tr	tr
Zircon	tr	tr	tr	tr	tr	tr	—	—	tr
Rutile	—	—	tr	—	—	—	—	—	—
Carbonate	—	—	—	—	—	tr	—	—	—
Magnetite-ilmenite	—	tr	0.5	tr	tr	tr	tr	tr	tr
Pyrite	—	—	—	—	—	—	tr	—	—
Per cent of anorthite in plagioclase	42	56	18	26	23	41	10	15	0
Grain size of phenocrysts (mm)	3.0–5.0	5.0–7.0	—	4.0–6.0	—	4.0–6.0	0.5–1.0	—	1.0
Grain size of groundmass (mm)	0.5–1.0	1.0–2.0	0.7–1.0	1.0–2.0	1.0–2.0	0.8–1.0	0.06–0.12	0.1–0.3	0.1–0.2
Texture	Por* Gr′	Por Gr	H″ M**	H S***	H M	H S	H S	H	H S

* Por—Porphyritic.
′ Gr—Granoblastic.
″ H—Hypidiomorphic.
** M—Mortar.
*** S—Subporphyritic.

Amphibolite sills and dikes.
1. Porphyritic amphibolite.
2. Porphyritic amphibolite

Granodiorite and quartz diorite stocks and sills.
3. Granodiorite.
4. Granodiorite.
5. Quartz diorite.
6. Quartz diorite.
7. Aplite.
8. Quartz diorite.
9. Soda granite, albitized.

general, the foliation parallels that of the country rock and, at least in some places, the contacts of the intrusives. The smaller bodies tend to be better foliated than the larger bodies. The largest pluton of this group covers about one-seventh of a square mile in the north-central part of the area and extends northward into the Bellows Falls quadrangle.

Most of these rocks are light-gray to medium-gray and medium-grained (Pl. 7, fig. 2), but fine-grained and even aplitic types are present. Most of them are equigranular, but some are subporphyritic. The essential minerals are plagioclase, quartz, potash feldspar, biotite, and muscovite. The plagioclase ranges from An_6 in some specimens to An_{41} in others. Apatite, zircon, sphene, hornblende, and garnet are accessory minerals. The quartz in some of these

bodies is highly granulated. Much of the quartz shows mortar and sutured texture, and all of it shows highly undulatory extinction. Very little of the feldspar is granulated. The average grain size of these rocks is about 1 to 2 millimeters; phenocrysts of quartz and feldspar reach a maximum length of about 6 millimeters. Aplitic varieties of these rocks tend to carry more sodic plagioclase than the coarser-grained rocks and contain no biotite. Estimated modes are given in Table 10, columns 3 to 9.

Intrusives of this group occur in the Orfordville, Partridge, Clough, and Littleton formations; hence they are younger than the lower Devonian Littleton formation. Some of these bodies crosscut the foliation of the country rock and contain folded inclusions of the country rock. Such inclusions are abundant around the borders of the small stock of quartz diorite 0.7 mile northwest of the village of Westmoreland. Because they contain inclusions of folded country rock, these bodies must have been intruded either late in the period of orogeny or after the orogeny. The mineralogy and structure of these intrusives indicate that they belong to the New Hampshire magma series. They are similar to the Haverhill granodiorite of the Mt. Cube quadrangle (Hadley, 1942).

A sill of highly sheared granitic rock about 1500 feet long and 200 feet wide intrudes the Orfordville formation immediately west of the Northey Hill thrust about 3 miles north of the city of Brattleboro. The foliation in the sill and in the surrounding schist strikes N. 15° E. to N. 25° E. and is vertical or dips steeply east or west. Similar highly sheared sills are found along, or near, the fault in other places. Although these sills show an exceptional amount of shearing, they are believed to belong to the New Hampshire magma series. In many places along the Northey Hill fault, the metamorphosed sedimentary rocks show a slip cleavage superimposed on the earlier flow cleavage; thus it appears that orogenic forces persisted longer in this structural belt than elsewhere. The shearing shown by the intrusives along this belt may be the result of these late stresses, whereas related intrusives in other parts of the area where these late stresses were not active are not sheared. Differences in amount of shearing in these bodies might also be accounted for by assuming that the sheared bodies are slightly older than the nonsheared.

Kinsman quartz monzonite.—A pluton of Kinsman quartz monzonite, in part porphyritic occupies about 50 square miles of the central part of the area. A body of schist of the Littleton formation about 3 miles long in a north-south direction and 1 mile wide is present in the east-central part of the pluton (Pl. 1). Inclusions of schist are so abundant in the quartz monzonite around this body, and sills and dikes of quartz monzonite so abundant in the body of schist, that in some cases it was difficult to decide whether a specific outcrop should be mapped as Kinsman quartz monzonite or as part of the Littleton formation. The area containing abundant inclusions is shown on the map (Pl. 1) by means of an overprint. Most of the inclusions are oriented parallel to the foliation of the enclosing Kinsman quartz monzonite. The inclusions range in size up to several tens of feet in length and thickness. Occasional small inclusions of schist are present throughout the Kinsman (Pl. 8, fig. 1).

The Kinsman shows good foliation near its contact with the Littleton formation, but throughout most of the interior of the body foliation is weak. Phenocrysts of feldspar are very sparse or absent along the western margin of the pluton but are abundant in most places in the interior of the body. The two long axes of most of the phenocrysts in foliated parts of the pluton parallel the plane of foliation (Pl. 8, fig. 1). Lineation produced by alignment of phenocrysts is rare.

In hand specimen, most of the Kinsman quartz monzonite is a medium-grained to coarse-grained light-gray porphyritic rock containing large tabular phenocrysts of white feldspar (Pl. 7, fig. 3). In most places, the phenocrysts are microcline or microperthite showing Carlsbad twinning; elsewhere they are plagioclase (An_8 to An_{37}). The largest phenocryst is about 40 by 25 by 15 millimeters. The percentage of phenocrysts is highly variable; they average about 10 per cent of the rock and range to a maximum of about 25 per cent. Owing to variation in the amount and composition of the phenocrysts, parts of the Kinsman body are petrographically granodiorite, other parts are

TABLE 11.—ESTIMATED MODES OF NEW HAMPSHIRE MAGMA SERIES

	1	2	3	4	5	6	7
Phenocrysts							
Potash feldspar	—	—	—	15	1	—	—
Plagioclase	—	10″	10	—	1	1	2
Per cent of anorthite in plagioclase	—	5	37	—	15	45	22
Size of phenocrysts in mm.	—	25.0	15.0	20.0	1.0–3.0	2.0–3.0	2.0–2.5
Groundmass							
Quartz	33	32	28	33	32	15	33
Plagioclase	50	38	30	30	6	56	45
Potash-feldspar	tr	—	—	tr	57	—	16
Biotite	12	10	18	12	2	20	2
Chlorite	—	tr	—	—	—	—	tr
Muscovite	5	7	11	10	tr	—	2
Sillimanite	tr	3	3	—	—	—	—
Garnet	tr	tr	tr	—	—	—	—
Hornblende	—	—	—	—	—	4	—
Zircon	tr	tr	tr	tr	tr	tr	tr
Apatite	tr	tr	tr	tr	tr	1	tr
Sphene	—	—	tr	—	—	3	—
Rutile	tr	—	—	tr	—	—	—
Tourmaline	—	—	—	—	1	—	—
Epidote	—	—	—	—	—	—	—
Allanite	—	—	—	—	—	tr	—
Magnetite	tr	tr	tr	tr	tr	tr	—
Pyrite	—	—	—	tr	—	tr	—
Per cent of anorthite in plagioclase	23	26	37	32	15	45	22
Size of groundmass (mm)	1.0–3.0	0.5–2.0	1.0–2.0	1.0–3.0	0.2–0.3	0.5–1.0	0.7–1.0
Texture	A* M‴	H** P***	H P	H P	H S″″	G″ S	H S

″—Intergrowth of 65% albite, 35% oligoclase.
* A—Augen.
‴ M—Mortar.
** H—Hypidiomorphic.
*** P—Porphyritic.
″″ S—Subporphyritic.
″ G—Gneissic.

Kinsman quartz monzonite.
1. Well-foliated quartz diorite.
2. Porphyritic quartz diorite.
3. Porphyritic quartz diorite.
4. Porphyritic quartz monzonite.
5. Aplite.

Spaulding quartz diorite.
6. Quartz diorite.

Concord granite.
7. Granodiorite.

quartz diorite. The different types are very similar in hand specimen.

The groundmass is chiefly feldspar, quartz, biotite and muscovite. Biotite is normally more abundant than muscovite and constitutes 18 or 20 per cent of the rock in a few places. Small pink garnets are not unusual, though never abundant. Sillimanite is conspicuous in many places and occurs as elongate grains up to 20 or 25 millimeters in length. It is more abundant where inclusions of schist of the Littleton formation are numerous, or close to the contact

with the Littleton formation, though it is not confined to such places. The distribution of the sillimanite indicates that the alumina was derived from inclusions of the Littleton formation. A specimen of quartz monzonite from Fullam Pond, near the center of the pluton, contains about 3 per cent sillimanite.

Microscopic study shows that the Kinsman quartz monzonite has a hypidiomorphic granular texture, with a tendency toward mortar and sutured texture. Essential minerals are oligoclase or andesine, quartz, potash feldspar, biotite, and muscovite. The plagioclase (An_{28} to An_{37} in different specimens) is present as homogeneous grains. Accessory minerals include garnet, sillimanite, zircon, apatite, pyrite, magnetite, sphene, and rutile. Microcline phenocrysts are ungranulated; most of the plagioclase phenocrysts have been granulated to fragments 0.1 to 0.3 millimeter in diameter. Most of the quartz occurs in large lens-shaped aggregates, but some occurs as interstitial grains or in myrmekitic or graphic intergrowth with plagioclase. The lens-shaped aggregates, up to 7 millimeters long and 4 millimeters wide, are probably single grains that have been crushed and now consist of 10 to 20 independently oriented granules. Most of the quartz shows strain shadows. Some of the quartz contains hairlike needles of sillimanite. The plagioclase of the groundmass is ungranulated and most of it is unaltered; albite twinning is well developed. Most of the fine-grained sillimanite is associated with and replaces muscovite. Zircon is an abundant accessory and produces prominent pleochroic halos in the biotite, with which it is associated. Estimated modes of the Kinsman quartz monzonite are given in Table 11.

Sills, dikes, and small irregular bodies of aplite or medium-grained white nonporphyritic muscovite-rich granitic rocks related to the Kinsman are abundant in many places along the contacts of the main intrusive, and locally these bodies cut the Littleton formation.

As shown by relationships along the contact between the two rocks, the Kinsman quartz monzonite is intrusive into, and younger than, the lower Devonian Littleton formation.

Spaulding quartz diorite.—The name Spaulding quartz diorite has been proposed by Katharine Fowler-Billings (1949) for a group of plutonic rocks of the New Hampshire magma series. The name was taken from Spaulding Hill, in the central part of the Monadnock quadrangle, where one of these plutons is exposed. A body of similar rock, about 1000 feet wide and 2200 feet long, is present 1 mile east of the city of Keene. Small stringers of the quartz diorite are also present at several places north of the main body.

In the Keene-Brattleboro area, the Spaulding quartz diorite is a foliated medium-gray to dark-gray medium-grained subporphyritic rock (Pl. 7, fig. 4). The color differs with the biotite content, which ranges from 15 to 25 per cent. Flakes of biotite, as much as 15 millimeters in diameter, oriented across the foliation are present in this rock in some places and indicate that recrystallization persisted after deforming stresses ceased. Sphene, an abundant accessory mineral, is easily identified megascopically. The sparse phenocrysts of plagioclase occur as grains 2 or 3 millimeters in diameter. As shown in Table 11, column 6, the average grain size of the groundmass is 0.5 to 1.0 millimeter; andesine (An_{45}), quartz, and biotite are the essential minerals; and sphene, apatite, hornblende, allanite, and pyrite are accessory minerals. Euhedral grains of zoned and twinned allanite are common. Most of them occur as inclusions in the biotite, where they display prominent pleochroic halos. The minerals present in this rock show almost no granulation.

In general, the foliation of the Spaulding quartz diorite parallels that of the surrounding country rock.

The Spaulding quartz diorite intrudes schists of the lower Devonian Littleton formation. In a few places, the quartz diorite contains thin sills of fine-grained light-colored granitic rock that closely resemble the Concord granite, and, at one place on the southeastern side of Beech Hill, this light-colored granitic rock contains inclusions of the quartz diorite.

Concord granite.—A body of Concord granite about 5 miles long and half a mile wide, elongate in a north-south direction parallel to the regional trend of the country rock, occupies the northeastern corner of the area and extends into neighboring quadrangles. Numerous thin sills and pods of this granite are found in the surrounding schists of the Littleton formation. In

hand specimen, the rock in this body closely resembles the Concord granite of surrounding areas. Although the one thin section from this body that was examined proved to be granodiorite (Table 11, column 7) the name Concord granite has been retained.

The Concord granite, in the Keene-Brattleboro area, is a light-colored fine-grained to medium-grained binary granodiorite with very weak to medium foliation. The micas occur as small flakes, and biotite is more abundant than muscovite in most of the rock. The granodiorite has not been crushed or granulated; the foliation is produced solely by parallel orientation or mica flakes. Small pink garnets are scattered throughout the rock.

Microscopic study shows that the rock has a hypidiomorphic subporphyritic texture, with a few 2.0- to 2.5-millimeter phenocrysts of plagioclase. The plagioclase of the groundmass (An_{22}) shows some myrmekitic intergrowth with quartz along grain boundaries. Quartz shows little undulatory extinction and no granulation. The potash feldspar is microcline with grid twinning.

The Concord granite intrudes the lower Devonian Littleton formation. Numerous sills and pods of Concord granite can be seen in the Littleton formation just west of the Dartmouth College Road near the northern edge of the Keene quadrangle.

Relative ages of the members.—The members of the New Hampshire magma series in the Keene-Brattleboro area, except for the amphibolite sills and dikes, are intrusive into the lower Devonian Littleton formation. The granodiorite and quartz diorite stocks and sills also intrude beds stratigraphically below the Littleton. Because all the members of the New Hampshire magma series show at least some granulation or foliation, they must have been intruded before the end of the period of deformation. Furthermore, the country rock was folded before the granodiorite and quartz diorite stocks and sills and the Kinsman quartz monzonite were intruded, for these two members contain inclusions of folded country rock.

The amphibolite sills and dikes of the New Hampshire magma series intrude gneiss of the Oliverian magma series. The amphibolite shows chilled borders against the gneiss, hence the country rock must have been relatively cool when the sills and dikes were intruded. The amphibolite, because it is foliated and contains granulated phenocrysts, was intruded before the end of the period of deformation. It is generally considered that early members of a magma series are the more mafic, whereas later members are more felsic. The amphibolite is probably an early member of the New Hampshire magma series rather than a late member of the Oliverian magma series.

TABLE 12.—RELATIVE AGES OF THE MEMBERS OF THE NEW HAMPSHIRE MAGMA SERIES, KEENE-BRATTLEBORO AREA

Member	Relations
Concord granite	Cuts Spaulding; contains inclusions of Spaulding. In other areas cuts Kinsman.
Spaulding quartz diorite	Cut by Concord; less deformed than Kinsman; massive in large bodies in other areas.
Kinsman quartz monzonite	More foliated and granulated than Spaulding; in other areas is cut by Spaulding.
Granodiorite and quartz diorite stocks and sills	More granulated than Kinsman, less recrystallized than amphibolite
Amphibolite sills and dikes	Cut Oliverian magma series, but more completely recrystallized than granodiorite and quartz diorite, well foliated.

The granodiorite and quartz diorite stocks and sills are not in contact with any of the other members of the New Hampshire magma series. These stocks and sills, in places, show fair foliation around their margins, but foliation decreases in intensity toward the centers of the bodies. The smaller bodies show better foliation than the larger ones; sills show better foliation than the cross-cutting stocks. The quartz in most of these bodies is granulated and is extremely so in some of the bodies. Mortar and sutured texture in the quartz is common; more-

over, all the quartz shows highly undulatory extinction. Although foliation is about as well developed in the Kinsman quartz monzonite as it is in the rocks of these stocks and sills, the Kinsman shows less granulation. The degree of granulation indicates that the stocks and sills are slightly older than the Kinsman quartz monzonite.

The Kinsman quartz monzonite is well foliated around the borders of the body; the interior of the body is locally well foliated. Bent grains of mica, bent plagioclase twin lamellae, and granulated quartz showing sutured and mortar textures indicate the amount of deformation to which the Kinsman was subjected. The Kinsman shows greater effects of deformation than does the Spaulding quartz diorite or the Concord granite and is assumed to be older than they are. Katharine Fowler-Billings (1949), from structural evidence in the adjacent Monadnock quadrangle, has concluded that the Kinsman is older than the Spaulding.

The Spaulding quartz diorite, in the Keene-Brattleboro area, shows fairly well developed foliation. However, the rock shows almost no granulation, and some of the coarse biotite flakes lie across the foliation. The foliation in some of the larger bodies of Spaulding in the adjacent Monadnock quadrangle (Katharine Fowler-Billings, 1949) is weak or absent. It appears that the Spaulding has been less deformed than the Kinsman, and hence is younger. The Spaulding is older than the Concord granite, for thin sills and stringers of Concord type granite cut the Spaulding.

The Concord granite is the youngest member of the New Hampshire magma series in the Keene-Brattleboro area. The Concord locally shows fair foliation, though in most places foliation is very weak. This rock shows no granulation.

Table 12 summarizes the relative ages of the members of the New Hampshire magma series in the Keene-Brattleboro area.

Age of the Intrusive Igneous Rocks

In the Keene-Brattleboro area, sills and dikes of the Oliverian magma series cut the Ammonoosuc volcanics. Inclusions of the volcanics are locally abundant near the margins of the plutons of the Oliverian magma series.

At the following two localities in the Mascoma quadrangle, the middle Silurian Clough formation has been feldspathized where it is in contact with granite of the Oliverian magma series (personal communication from C. A. Chapman and R. W. Chapman to M. P. Billings, (1946):

(1) in the brook along the trail on the western side of South Peak, at an altitude of 1600 feet to 1750 feet.

(2) on the two small knobs one-quarter mile west of the S in Moose Mountain.

In the Bellows Falls quadrangle, Kruger (1946, p. 181) states that "on the southeast side of the Alstead dome the gneiss (Oliverian magma series) intrudes schists, some of which belong to the Partridge formation and some of which may belong to the Littleton formation (lower Devonian)."

Intrusive relations between the gneiss and the surrounding meta-sediments thus show that the Oliverian magma series is younger than middle Silurian, and possibly younger than lower Devonian. Sills and dikes of strongly foliated amphibolite intrude gneiss of the Oliverian magma series in the Keene-Brattleboro area; the intrusion of the gneiss was therefore pre-folding.

The Oliverian magma series has been assigned to the middle Devonian (?).

The New Hampshire magma series is younger than the lower Devonian, for granodiorite and quartz diorite stocks and sills, Kinsman quartz monzonite, Spaulding quartz diorite, and Concord granite intrude the Littleton formation. Structural evidence indicates that the New Hampshire magma series is younger than the Oliverian magma series.

In central New Hampshire, intrusive relations show that the White Mountain magma series is younger than both the Oliverian magma series (Chapman, Billings, and Chapman, 1944, p. 510) and the New Hampshire magma series (Williams and Billings, 1938). The White Mountain magma series has been correlated with the Quincy granite of Massachusetts (Williams and Billings, 1938, p. 1025), which is pre-Pennsylvanian (Emerson, 1917, p. 188). Thus the New Hampshire magma series and the White Mountain magma series are younger than lower Devonian and older than Pennsylvanian. The New Hampshire magma series has been assigned

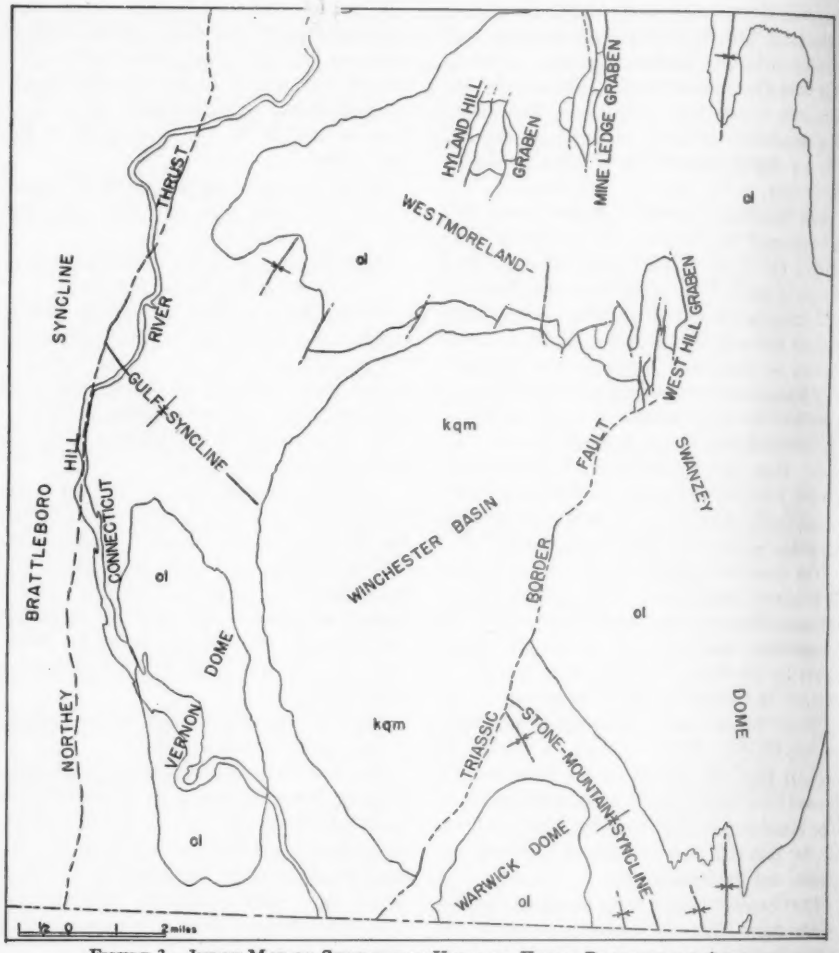

FIGURE 3.—INDEX MAP OF STRUCTURAL UNITS OF KEENE-BRATTLEBORO AREA

to the late Devonian(?), and the White Mountain magma series to the Mississippian(?).

GENERAL STRUCTURAL FEATURES

The rocks of the Keene-Brattleboro area show a great variety of structural features. Some of these are regional and extend for a great distance along the trend of the structure, others are local. The largest structural units in this area are the Bronson Hill anticline, the Northey Hill thrust, and the Brattleboro syncline. The Northey Hill thrust extends from a point about 4 miles east of the northwestern corner of the area to a point about 1 mile east of the southwestern corner of the area. The Bronson Hill anticline, modified by conspicuous subsidiary folds with amplitudes as much as several thousands of feet, occupies the entire area east of the thrust. The Triassic border fault enters the area just west of Hogback Mountain, trends northeasterly along the Bronson Hill anticline, and apparently dies out near Keene. Numerous small high-angle normal faults are present on the Bronson Hill anticline, and in three localities graben form very striking

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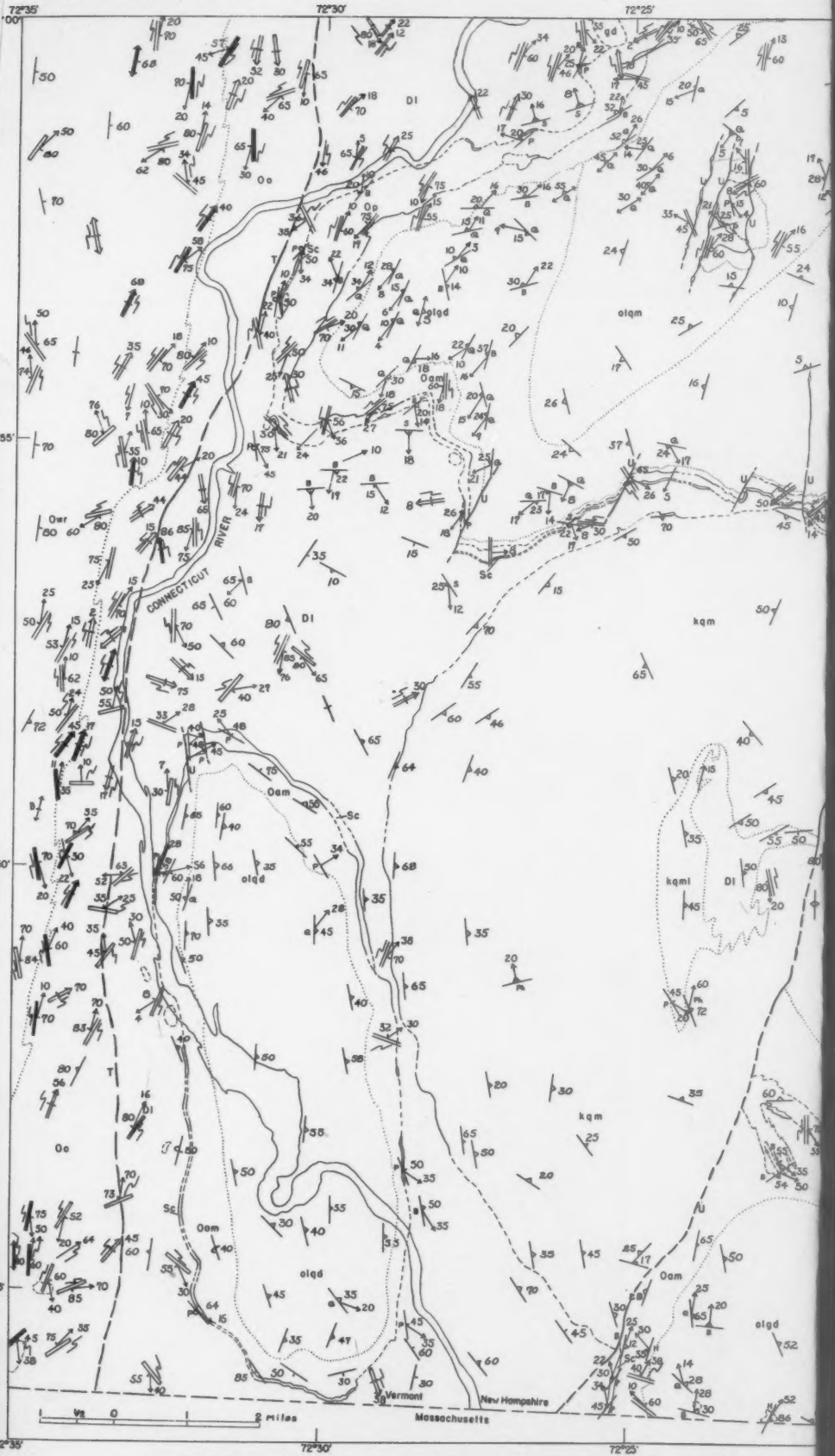
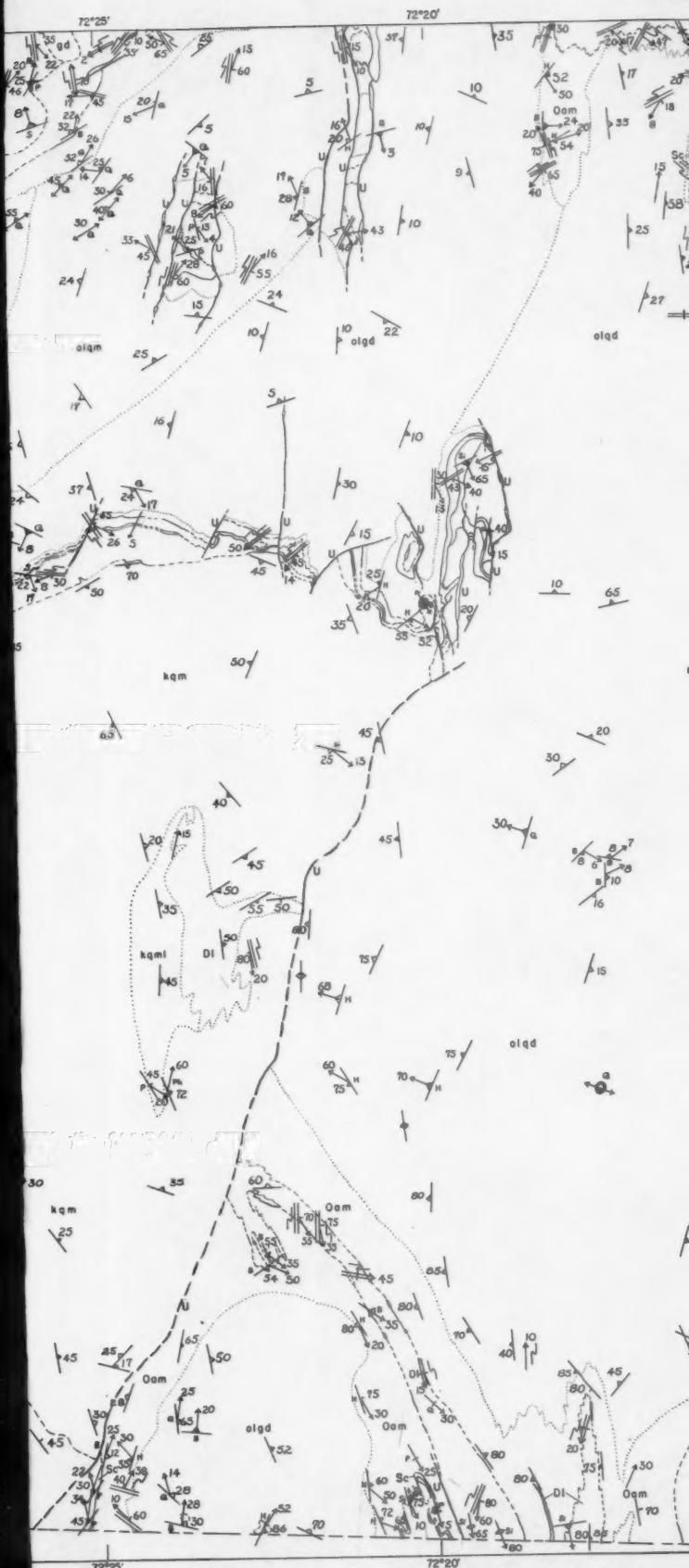


TECTONIC MAP OF KEENE-BRATTLEBORO AREA, NEW



TLEBORO AREA, NEW HAMPSHIRE-VERMONT

Moore, Pl. 2

LEGEND

Structural Symbols

50/ Bedding

Bedding where vertical

60/ Bedding where overturned

75/ Foliation

Foliation where horizontal

Foliation where vertical

20/ Lineation - strike of horizontal projection of linear element and value of plunge

Horizontal lineation

Symbols for lineation are used in conjunction with those for bedding or foliation

Nature of linear element:

P - Pebbles Ph - Phenocrysts
Q - Quartz
B - Biotite
M - Muscovite
H - Hornblende
S - Staurolite
Si - Sillimanite

Older Folds

30/50 Fold - strike and dip of axial plane and horizontal projection of axis and value of plunge

40 Fold - axis plunges down the dip of the axial plane

20 Fold - vertical axial plane and value of plunge of axis

75 Fold - strike and dip of axial plane and horizontal axis

Fold - vertical axial plane and horizontal axis

Younger Folds

55/70 Symbols as above

Unclassified Folds

55/70 Symbols as above

15/ Fold - strike of horizontal projection of axis and value of plunge

Plan of folds

Formation symbols, Contacts, Faults — Same as on geological map

ONT

structural features. The Brattleboro syncline occupies the area west of the Northey Hill thrust. The eastern limb of the syncline has been cut out by the thrust. Numerous intrusive bodies which differ greatly in size and shape are present in the area. The general trend of the folds and the larger plutons is north; the trend of most of the normal faults is N. to N. 30° E.

Foliation is well developed in all the metamorphic rocks of the area. In most of the metamorphosed stratified rocks east of the Northey Hill thrust foliation and bedding are parallel, whereas in most of the rocks west of the Northey Hill thrust foliation is parallel to the axial planes of folds. Minor folds of two ages are present west of the Northey Hill thrust and in a few localities just east of the thrust. Most of the older minor folds are isoclinal, or nearly so, and have a well-developed axial-plane flow cleavage. The younger minor folds are more open and have an incipient to strong axial-plane slip cleavage. The incipient slip cleavage produces crinkling, the trace of which on flow cleavage or bedding-plane surfaces gives a strong lineation. Oriented minerals, or mineral aggregates, and elongated pebbles produce a marked lineation in the rocks in many localities.

The principal structural data of the Keene-Brattleboro area are recorded on Plates 1–3.

Major Folds, Minor Folds, and Lineation

General Statement

The folds of greatest magnitude are the Bronson Hill anticline and the Brattleboro syncline. Plutons of the Oliverian magma series and the New Hampshire magma series are present along the Bronson Hill anticline.

Bronson Hill Anticline

General statement.—The Bronson Hill anticline has now been mapped from the Mount Washington quadrangle (Fig. 1) southwestward into Massachusetts, a distance of about 140 miles. Numerous elongate domes, each of which has a core of plutonic rock belonging to the Oliverian magma series and a surrounding envelope of metamorphic rocks, are present along the crest of the Bronson Hill anticline. From the central part of the Mt. Cube quadrangle (Fig. 1) southward, the long axis of each dome trends about north-south, whereas the anticline trends about N. 15° E. The domes are thus en echelon, and each dome, toward the south, is progressively farther west. Northward from the Mt. Cube quadrangle the axis of the anticline curves more toward the northeast, and the long axis of each dome is about parallel to the trend of the anticline.

The Bronson Hill anticline, in the Keene-Brattleboro area, is about 15 miles wide and occupies the entire area east of the Northey Hill thrust. Three elongate domes and numerous subsidiary folds are present on the anticline in this area. The Westmoreland-Swanzey dome, which extends a short distance northward into the Bellows Falls quadrangle, occupies the northern and eastern parts of the area and is complicated by subsidiary folds. The northern end of the Warwick dome occupies the south-central part of the area, and it also is complicated by subsidiary folds. The Vernon dome is present in the southwestern part of the area. The Stone Mountain syncline, a tight narrow fold, is present between the southern part of the Westmoreland-Swanzey dome and the Warwick dome. The Gulf syncline is present between the northwestern part of the Westmoreland-Swanzey dome and the Vernon dome. The Winchester basin, in the central part of the area, near the center of the Bronson Hill anticline, is occupied by a thick sheet of Kinsman quartz monzonite.

Westmoreland-Swanzey dome.—The Westmoreland-Swanzey dome, in the northern and eastern parts of the Keene quadrangle, includes a core of gneiss of the Oliverian magma series (the Westmoreland-Swanzey pluton) and the surrounding envelope of metamorphosed sedimentary and volcanic rocks. The envelope of stratified rocks includes the Ammonoosuc, Partridge, Clough, and Littleton formations.

The stratified rocks are absent on the western side of the eastern lobe of the dome, where gneiss of the Oliverian magma series is in contact with Kinsman quartz monzonite along the Triassic border fault. The stratified rocks above the horizon intruded by the Oliverian magma series have been removed from the eastern, upthrown side of the fault by erosion. On the western, downthrown side of the fault, the stratified rocks between the horizons intruded by the Oliverian magma series and the Kins-

man quartz monzonite underlie the quartz monzonite.

The Partridge formation is absent along most of the southern border of the eastern lobe of the dome; the Clough quartzite is discontinuous around the entire dome.

Foliation is perceivable in nearly every outcrop of the Westmoreland-Swanzey pluton, but it is best developed around the border of the body. The foliation in the interior of the plutons of the Oliverian magma series is not so well developed as the structure sections on Plate 1 indicate. In most places along the border of the body, foliation in the pluton dips radially outward from the center of the dome. Both foliation and bedding, mutually parallel, are well developed in the stratified rocks of the dome. In plan, the foliation and bedding in the Ammonoosuc volcanics, the foliation in the Oliverian magma series, and the contact between these two rocks are mutually parallel.

The contact between the Ammonoosuc volcanics and the gneiss of the Oliverian magma series was observed in a number of localities. On the south side of the 1000-foot hill, 0.25 mile north of the Richmond Trout Rearing Station, in the eastern part of Richmond, the contact is exposed in a cliff face about 25 feet high. The contact at this locality is parallel to both the foliation and bedding in the Ammonoosuc volcanics and to the foliation in the Oliverian magma series. This same relation was observed in the other exposures of the contact, most of which were in faces only 5 to 10 feet high. The contact is presumed to be parallel to the foliation and bedding in the adjacent rocks elsewhere.

The average outward dip of the foliation along the border of the northwestern lobe of the Westmoreland-Swanzey pluton, from Gilboa southwestward across Cass Hill, Bald Hill, and Streeter Hill, thence eastward through Spofford to West Hill, is 25° to 40°. At the extreme western end of the northwestern lobe of the dome, between Cass Hill and Streeter Hill, the foliation in the gneiss of the Oliverian magma series, in the Ammonoosuc volcanics, and in the lower part of the Partridge formation dips outward from the center of the dome. Toward the west, the dips steepen and become overturned, until the upper part of the Partridge, the Clough, and the lower part of the Littleton dip eastward at angles of 45° to 80° (Pl. 1).

The dip of the foliation around the border of the eastern lobe of the Westmoreland-Swanzey pluton, in general, is steeper than that around the northwestern lobe (Pl. 2). Along the northeastern side of the eastern lobe, north of the city of Keene, the dip of the foliation in the gneiss, and in the overlying Ammonoosuc volcanics, ranges from 30° to 65° E. At the northwestern extremity of the eastern lobe, 3 miles N. 20° W. of the city of Keene, around Goose Pond, the structure is overturned, for both the foliation in the gneiss and the foliation and bedding in the overlying Ammonoosuc volcanics dip 30° E. to 65° E. Along the western side of the eastern lobe, between West Swanzey and Scott Mountain, dips average 50° to 80° W. Around the southern end of this lobe, foliation in the pluton dips outward at angles of 40° to 75°. The dip of the foliation over most of the interior of the Westmoreland-Swanzey pluton is low; dips range, in general, from zero to 30° and indicate a series of broad flexures.

Major folds and associated minor structures of the Westmoreland-Swanzey dome.—A tightly folded syncline of Ammonoosuc volcanics that forms Surry Mountain, 3 miles N. 20° W. of the city of Keene, extends southward from the Bellows Falls quadrangle for about 2¼ miles. This syncline is overturned toward the west. The axial plane of the syncline strikes about N. 5° E.; dips on both limbs of the fold range from 30° E. to 70° E. (Pls. 1, 3). Near the northern edge of the Keene quadrangle, the axial planes of minor folds on this syncline strike NNE., and have an average dip of about 75° E. The axes of most of the minor folds plunge 20° to 40° NNE. Foliation and plunges have been inverted at the southern end of the syncline, for there the foliation dips 30° to 50° S., and minor folds plunge 20° to 50° SW. Mineral lineation in the rocks of this syncline tends to strike across the trend of the fold.

A broad open syncline, flanked by anticlines on either side, is present on the southwestern side of the northwestern lobe of the dome, 1.5 miles northwest of Spofford Lake. The axial

plane of the syncline strikes N. 30° E. and dips steeply southeast; the axis plunges 10° to 20° SW., as shown by bedding on the nose of the fold. The dip of the bedding on the limbs of the fold is low.

A series of closely folded anticlines and synclines occurs at the southern end of the dome in the town of Richmond. The crest of a narrow anticline follows the valley of Sprague Brook; two synclines and two anticlines are present to the west of the brook. The axes of the two synclines are respectively 0.5 mile and 0.95 mile west of where Sprague Brook crosses the state line. The crest of the anticline along Sprague Brook is occupied by gneiss of the Oliverian magma series in some places, and by Ammonoosuc volcanics elsewhere. The two anticlines to the west expose Ammonoosuc volcanics, whereas schists of the Littleton formation occupy the troughs of the two synclines. The more easterly folds strike nearly north-south; the more westerly folds strike north near the southern edge of the quadrangle, but farther north they strike NNW. All the folds plunge gently south. The belt of schist in the core of the more easterly of the two synclines is about 2 miles long and 0.25 to 0.5 mile wide, and a lens of Clough quartzite is present on the northeastern side of the syncline. The band of schist in the western syncline is about 0.75 mile long and 200 to 400 feet wide. All these folds extend southward into the Mt. Grace quadrangle, Massachusetts. Bedding and foliation dip outward from the crest of the anticline along Sprague Brook at angles of 55° to 90°. Near the southern edge of the quadrangle, bedding and foliation on both limbs of the eastern syncline show vertical to steep easterly dips; while near the northern end of the syncline bedding and foliation dip 55° to 80° W. Along the southern edge of the quadrangle bedding and foliation in the western syncline dip 70° to 85° E.; at the northern end of the syncline they dip 75° to 85° W. The axial planes of these isoclinal folds show the same changes in dip as do the bedding and foliation.

The description of the next syncline to the west, the Stone Mountain syncline, is reserved for a later page.

The discussion of the Westmoreland-Swanzey dome has been confined to the broader structural features. It is now appropriate to consider the minor structural features of this dome and their relation to the major features.

The location and general attitude of the larger folds on the Bronson Hill anticline are clearly revealed by the areal distribution of formations and by the attitude of the bedding and foliation. It is not necessary to rely on the attitude and pattern of drag folds or the attitude of lineation to solve the structure of this part of the area. However, minor folds are abundant in the stratified rocks in many places; and lineation, shown by parallel orientation of stretched pebbles in conglomerate, by elongate mineral grains in the metamorphic rocks, and by quartz eyes and biotite splotches in the gneiss of the Oliverian magma series, is locally well developed. The area offers good opportunity to compare the attitude of the major folds with the attitude of minor structural features.

Point diagrams showing the attitude of axes of minor folds too small to show on the scale of the map, and the attitude of lineation, are presented on Plate 3. Each diagram includes all the data on minor fold axes and lineation that were observed in the field in the specific area covered by the diagram. Areas covered by individual diagrams were selected, insofar as possible, so they would not include parts of different structures. Conventional symbols are used on Plate 3 to show the attitude of minor folds and lineation in areas where few data on these features were obtained. The data are plotted on the upper hemisphere of a Schmidt equal-area net. In area 21, for example, the lineation plunges 14° to 28° in a direction ranging from N. 5° E. to N. 12° W.; and in area 22, the fold axes and lineation plunge 0° to 40° in a direction ranging from S. 15° E. to S. 45° E.

The attitude of the axes of the minor folds and the lineation is not necessarily the same as the attitude of the larger folds (Pl. 3). In some places the axes of minor folds and the lineation are not only parallel to each other but are also parallel to the axes of the larger folds, as in areas 3, 6, 15, 20, 21, and 22, In places, such as areas 4, 8, and 9, the axes of the minor folds roughly parallel the axes of the larger folds,

whereas the lineation plunges more northerly or more southerly. In places, as area 24, most of the lineation is about parallel to the axes of the larger folds, whereas the axes of the minor folds are not parallel to those of the larger folds. In places, as area 28, the trend of the lineation is almost at right angles to the trend of the larger folds; and in still other places, as areas 2 and 25, there appears to be no consistent relation between the larger folds and these minor structures. Locally, as in area 5, lineation shows concentrations in more than one direction.

Vernon dome and associated minor structures.—The Vernon dome, in the southwestern part of the area, is composed of a core of quartz diorite gneiss (the Vernon pluton) about 8 miles long and 2 miles wide, surrounded by concentric bands of metamorphosed sedimentary and volcanic rocks. The lowlands from Wantastiquet Mountain southward to the Vermont-Massachusetts state line are underlain by the quartz diorite; the surrounding hills are held up by the more resistant envelope of metamorphosed sediments and volcanics. Except where the Connecticut River crosses the dome, topography reveals very strikingly the distribution of the lithologic units.

The Partridge formation, present in the Westmoreland-Swanzey dome to the north, occurs in the Vernon dome only at the extreme northern end, where it is too thin to show on the scale of Plate 1. Conversely, the Clough quartzite is thicker and more persistent in the Vernon dome than in the Westmoreland-Swanzey dome.

Foliation in the gneiss parallels foliation and bedding in the surrounding metamorphic rocks. Available data indicate that the contact between the gneiss and the overlying Ammonoosuc volcanics is concordant. Foliation is well developed around the border of the quartz diorite but decreases in intensity toward the interior of the body.

The strike of the bedding and foliation in the stratified rocks swings around the northern end of the dome, where dips are toward the north. Along the northeastern side of the dome, foliation in the quartz diorite dips 30° E. to 50° E.; dips gradually steepen in the metamorphic rocks to the east. On the eastern and southeastern side of the dome, dips in both the quartz diorite and the stratified rocks range between 40° E. and 70° E. As the stratified beds swing around the southern end of the dome, the dips steepen and become vertical on the southwestern flank of the dome. The entire western half of the dome is overturned, for bedding in the metamorphic rocks and foliation in the quartz diorite dip 40° to 70° E.

At the northern end of the dome (area 15, Pl. 3) most of the minor folds and linear features plunge 30° to 60° N. 50° E. to N. 90° E. Because the dome is strongly overturned, the direction of plunge of the minor folds and lineation is not parallel to the axial trace of the dome. Along the northeastern side of the dome the minor folds and lineation plunge to the northeast; on the southeast side of the dome they plunge to the southeast. Near the southern end of the overturned western limb of the dome, the lineation of elongate pebbles plunges to the southeast, while near the northern end of the western limb similar lineation plunges to the northeast.

Warwick dome.—The northern part of the Warwick dome, with its core of granodiorite gneiss of the Oliverian magma series (the Warwick pluton) and the surrounding concentric belt of stratified rocks, occupies the south-central part of the area. In the Keene-Brattleboro area, the core of granodiorite underlies the lowlands south of the village of Winchester, and the metamorphosed sedimentary and volcanic rocks hold up the bordering hills. The body of granodiorite is 3 miles wide along the southern border of the area. Much of the northern and eastern parts of this body, along Sunny Valley, are poorly exposed.

In general, foliation in the granodiorite gneiss is parallel to the foliation and bedding in the overlying Ammonoosuc volcanics. Foliation and bedding on the eastern side of the dome dip 50° to 80° E. On the western side of the dome, the contact of the granodiorite with the volcanics shows considerable folding, and strikes and dips are irregular. At the Massachusetts-New Hampshire state line, foliation and bedding in the Ammonoosuc volcanics near the contact strike northerly and dip 45° to 70° E. (Pl. 1). Along the road 0.5 mile to the north, foliation near the contact dips 25° N. One mile north of the state line, foliation in the granodiorite and in the volcanics dips 30° to 60° W. From Manning Hill northward, foliation dips 45° to 75° E. The

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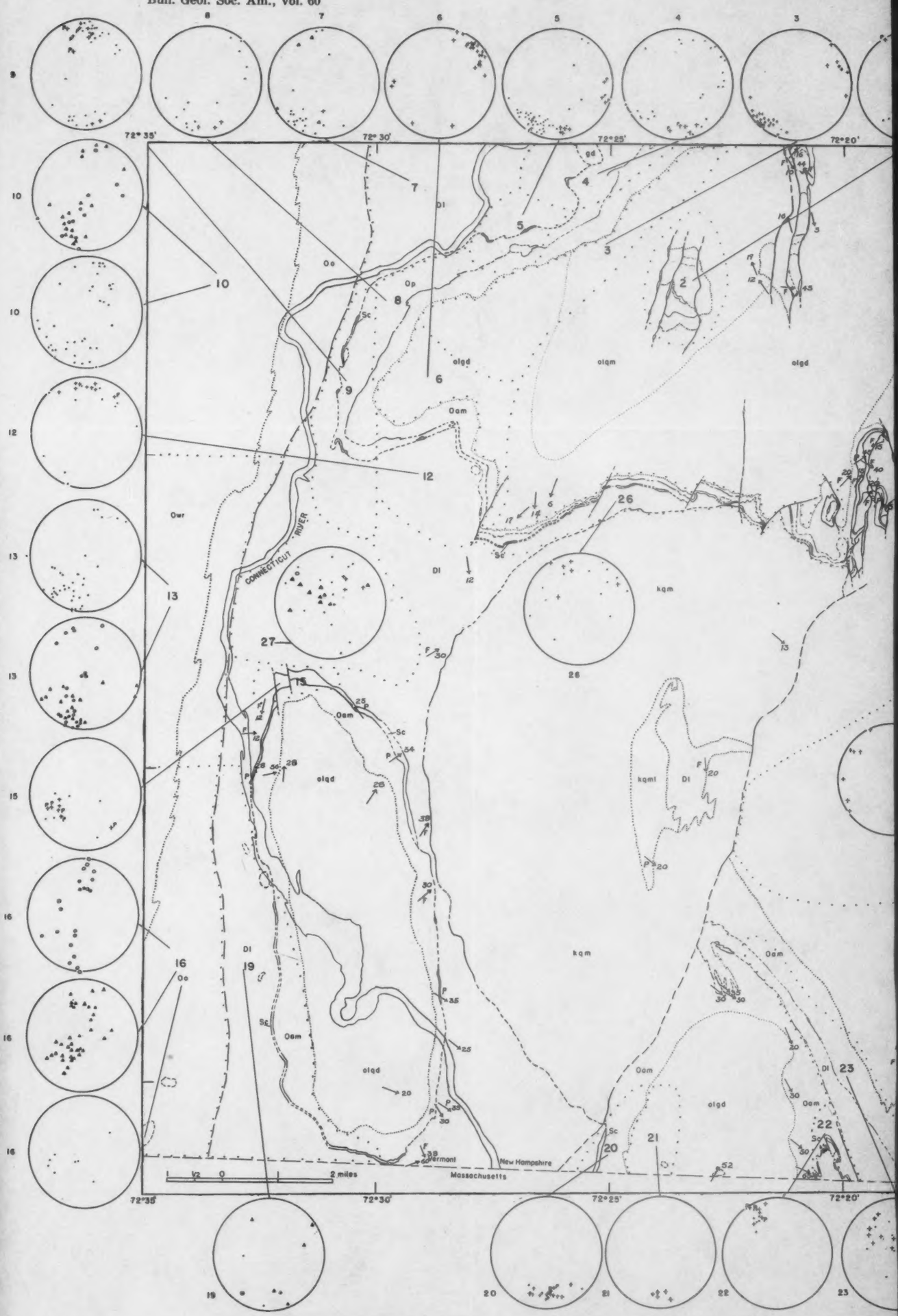
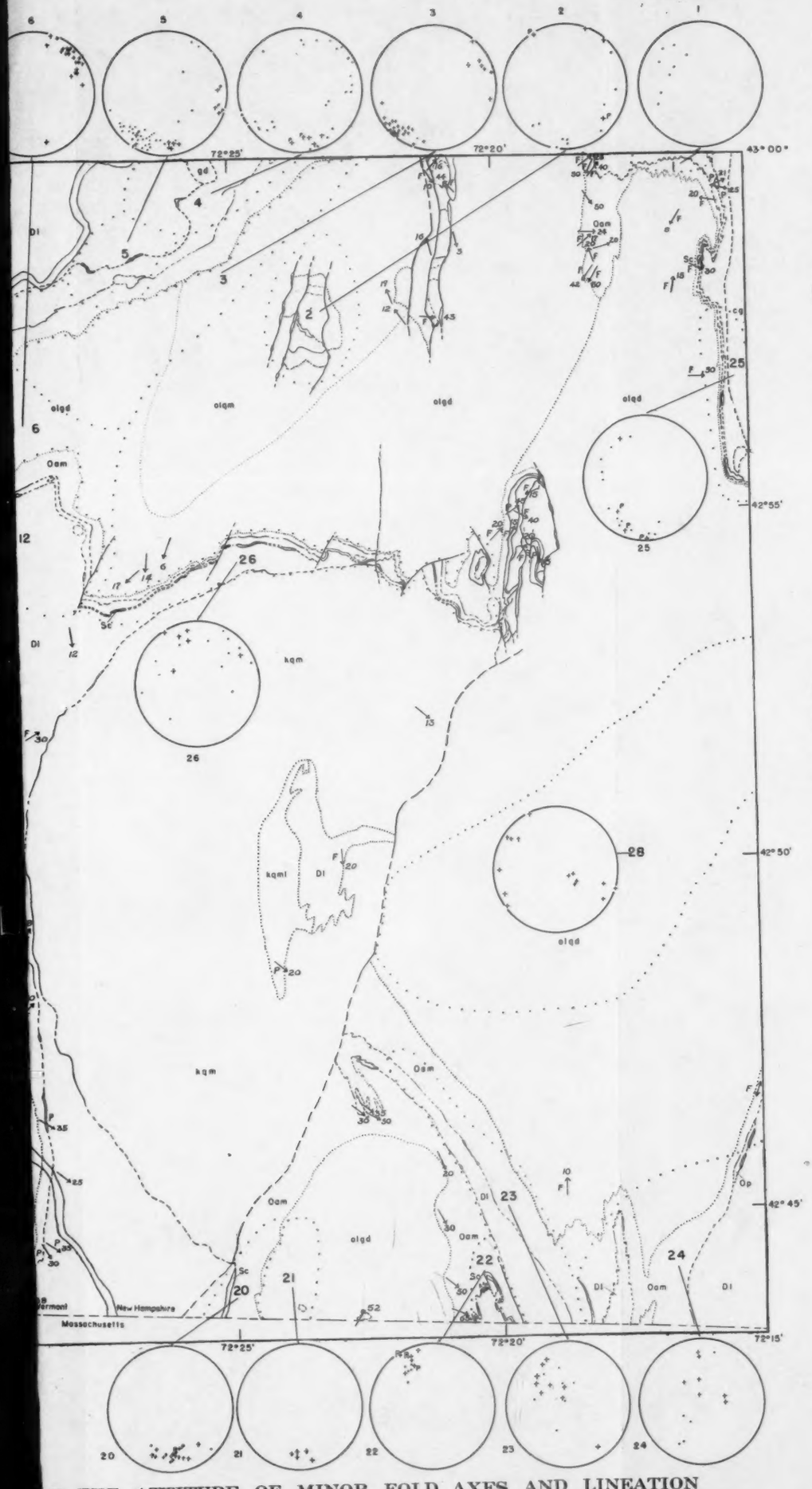


POINT DIAGRAMS OF THE ATTITUDE OF MINOR FOLD AXES AND



OF THE ATTITUDE OF MINOR FOLD AXES AND LINEATION

northern half of the western side of the dome is thus overturned toward the west. Bedding in the Ammonoosuc volcanics at the northern end of the dome dips N. or NE.

Shear zones in the granodiorite, along which foliation shows slight offset, in some places are filled with pegmatite that grades into the gneiss.

The contact between the granodiorite and the Ammonoosuc volcanics as exposed in a small fold along the road half a mile north of Hogback Mountain parallels the bedding in the volcanics, but the foliation in the granodiorite cuts across the contact at a high angle (Fig. 4B; Pl. 8, fig. 3). Thus, in this minor fold, the foliation in the granodiorite must be secondary. In another fold in the same road cut, the foliation in the volcanics wraps around the nose of the fold, parallel to the bedding of the volcanics (Fig. 4A; Pl. 8, fig. 4).

These exposures have great significance in the interpretation of the geologic history of the area. Inasmuch as the concordant contact between the Oliverian magma series and the Ammonoosuc volcanics is folded, the folding is younger than the intrusion of the Oliverian magma series.

Mineral lineation in the Ammonoosuc volcanics on the eastern side of the dome plunges 30°–65° SSE. to SE. (Pl. 3). Lineation is exceptionally well developed on the western side of the dome (areas 20 and 21, Pl. 3), where it plunges N. 20° E. to N. 20° W. at angles of 20° to 35°. Lineation is expressed by stretched pebbles in the Clough quartzite, by elongate minerals and fluting in the Ammonoosuc volcanics, and by quartz eyes and biotite clusters in the granodiorite of the Oliverian magma series.

Major folds and associated minor structures of the Warwick dome.—A tightly folded syncline containing a core of schist of the Littleton formation is present on the eastern flank of the Warwick dome. The core of schist, which extends northward from the Mt. Grace quadrangle, is about a mile long and about three-quarters of a mile in maximum width. The Clough quartzite displays very clearly the pattern of this fold (Pl. 1). Both limbs of the fold dip 60° E. to 90°. The western limb trends about N. 30° E., whereas the bedding and foliation strike N. 15°–30° W. Tight folding, and

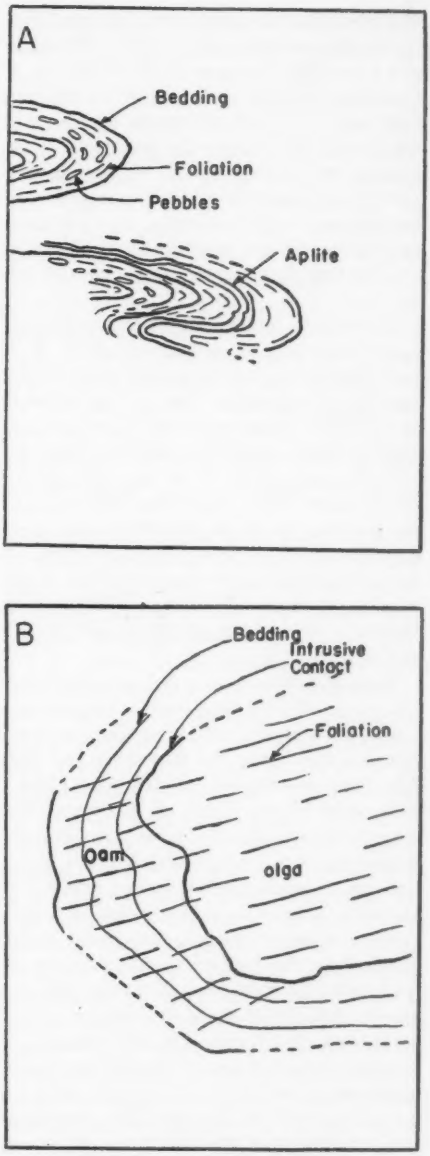


FIGURE 4.—SKETCH OF FOLDS

perhaps a good deal of slicing, explains the difference in the strike of the bedding and the trend of the limb of the fold. Near the northern end of the syncline, the Clough quartzite,

50–75 feet thick in this locality, has been piled up by tight folding, and probably slicing, to give a breadth of outcrop of 500 to 750 feet. A small body of schist, on the nose of the fold, has been almost squeezed off from the main body of the Littleton formation. The axial plane of the syncline strikes about N. 30° W. and dips 60° to 90° E.; dips tend to be slightly steeper toward the northern end of the syncline. Linear features and minor folds plunge 0° to 40° in directions ranging from S. 15° to S. 45° E. (area 22, Pl. 3).

A complexly folded syncline forms Meetinghouse Hill, near the northern end of the Warwick dome. This fold is on the strike of the syncline just described. The band of schist of the Littleton formation in the core of the syncline is about a mile long and a quarter of a mile wide and is cut off on the northwest by the Triassic border fault. The axial plane of the syncline trends about N. 30° W.; dips average 55°–70° E. on both limbs of the fold. A strong lineation near the southern end of the syncline has an average plunge of 30° to 50° in directions ranging from S. 40° E. to S. 75° E. (Pl. 3).

Stone Mountain syncline and associated minor structures.—The Stone Mountain syncline separates the southern end of the Westmoreland-Swanzey dome from the Warwick dome. This fold trends southeasterly from a point about 1 mile north of the village of Winchester and extends through Stone, Peaked, and Wheeler mountains, and Whipple Hill, where it leaves the area. A band of schist of the Littleton formation a quarter to three-quarters of a mile wide and about 6 miles long occupies the trough of the fold; Ammonoosuc volcanics occur on both limbs. The fold is cut off on the northwest by the Triassic border fault. Another normal fault, west of Whipple Hill, cuts out part of the western limb of the fold. Quartzite and conglomerate of the Littleton formation, having a maximum thickness of about 50 feet, are present near the trough of the fold in the vicinity of Whipple Hill. The quartzite on the eastern limb of the fold crosses the crest of Whipple Hill, where it is repeated by tight folding, or slicing, to give a breadth of outcrop of about 400 feet. The same quartzite occurs on the western limb of the fold, 1000 feet west of Whipple Hill. The axial plane of the fold strikes N. 20°–40° W. and dips 70°–90° E. Most of the linear features and minor folds plunge 20° to 65° in directions ranging from S. 10° E. to S. 65° E. (area 23, Pl. 3).

Gulf syncline and associated minor structures.—The Gulf syncline, which strikes across the western flank of the Bronson Hill anticline, separates the western end of the northwestern lobe of the Westmoreland-Swanzey dome from the Vernon dome. The axis of the fold extends about N. 60° W. from near Davis Hill to the Connecticut River, passing just north of the Gulf; thence it swings toward the north and is cut off by the Northey Hill fault southwest of Dummerston Station.

The Littleton formation, in this syncline, has a breadth of outcrop of about 4 miles; successively older formations, dipping away from the domes to the north and south, appear on both limbs. Southerly dips, away from the dome to the north, average 15° to 35°; they steepen to 75° to 90° near the axis of the syncline. Northerly dips, away from the Vernon dome, average 40° to 60°, steepening to 75° to 90° near the axis of the syncline. The axial plane of the syncline thus dips steeply to the south and is much nearer the Vernon pluton than the Westmoreland-Swanzey pluton.

Toward the center of the syncline the rocks display minor folds of two ages. One group of folds, probably older folds but shown on the diagram for area 27 (Pl. 3) as unclassified folds, are nearly isoclinal and have an axial-plane flow cleavage. The flow cleavage cuts across bedding on the noses of the folds, but on the limbs of the folds the bedding and the flow cleavage appear to be parallel. The axial planes of most of these folds strike approximately north-south, across the axis of the syncline, and dip steeply to the east or west; the axes plunge north or south. The younger folds are more open folds and have an incipient to strong axial-plane slip cleavage. The axial planes of most of the younger folds strike northwest, about parallel to the axis of the syncline, and are vertical or dip steeply south; the axes plunge southeast at gentle to steep angles. These relations suggest that the major syncline is itself a younger fold.

Toward the north and south, away from the

axis of the syncline, the Westmoreland-Swanzey dome and the Vernon dome show increasing control on the attitude of the minor structural features.

Winchester basin.—The Winchester basin is in the central part of the area and is almost surrounded by the three domes of the Keene-Brattleboro area. The basin is occupied by a pluton of Kinsman quartz monzonite, 11 miles long in a north-south direction and 6 miles wide, intrusive into schists of the Littleton formation. The eastern side of the pluton is cut off by the Triassic border fault. A body of schist of the Littleton formation, about 3 miles long and 1 mile wide, is present in the east-central part of the Kinsman quartz monzonite.

Along the borders of the pluton the foliation and bedding in the schist and the foliation in the quartz monzonite conform to the oval pattern of the intrusive and dip 45° to 80° toward its center. The contact was seen in a number of localities in faces up to 5 or 10 feet high. In these exposures the contact appears to parallel the foliation of the rocks on either side. The borders of the pluton are well foliated, but in most parts of the interior of the body foliation is weak, and dips are lower, averaging 30° to 45°. The foliation and bedding of the included body of schist also parallel the foliation in the surrounding Kinsman quartz monzonite. The dip of the foliation in and around this body of schist is highly variable, but most of the dips are relatively low.

Although the foliation in the schist of the Littleton formation is parallel to the contact of the Kinsman quartz monzonite in plan, and is locally so in section, the quartz monzonite does cut across the structure of the enclosing rock in some places, for the thickness of the Littleton formation present below the horizon intruded by the quartz monzonite is highly variable. Three miles east of Spofford Lake only 185± feet of the Littleton formation is present between the Kinsman quartz monzonite and the top of the Clough quartzite, whereas 1 mile east of Spofford Lake this interval is 1500± feet. On the western side of the pluton, 3.5 miles north of the village of Hinsdale, about 1100 feet of schist separates the Clough from the Kinsman quartz monzonite, and on Kilburn Brook, 1.5 miles north of the village of Hinsdale, this interval is about 400 feet. Discordance of the minor features of this pluton was recognized by Daly (1897, p. 717).

Dikes of Kinsman cut the schist along the western side of the pluton. Inclusions of schist in the Kinsman are numerous near the contact and are locally present in the interior of the body. Inclusions are so numerous around the large body of schist in the eastern part of the pluton that it is difficult to decide whether certain outcrops should be mapped as Kinsman full of inclusions, or as schist containing abundant sills of Kinsman. Most of the inclusions are small, are shaped like a slightly elongated disc, and parallel the foliation of the enclosing rock. Some of them, for example one three-quarters of a mile north of Hale Pond, on the east side of the road, are tens of feet long and are not parallel to the foliation of the enclosing rock.

The foliation of the Kinsman quartz monzonite is produced by parallel arrangement of muscovite and biotite flakes and, to a lesser extent, by parallel orientation of feldspar phenocrysts. Linear parallelism is very rare.

The foliation in the Kinsman quartz monzonite is believed to be a primary flow structure due to the forceful injection of the magma, shortly before the end of the orogeny, for the following reasons: (1) Most of the abundant inclusions are oriented parallel to the foliation of the enclosing quartz monzonite. (2) Most of the feldspar phenocrysts, believed to be intratelluric, are not highly granulated, and in most places they lie in the plane of foliation. (3) The margins of the pluton, in general, are better foliated than is the interior. (4) The inclusions were foliated before incorporation into the quartz monzonite, for the foliation of some inclusions of schist is not parallel to the foliation in the quartz monzonite. These discordant inclusions show no trace of foliation parallel to that of the quartz monzonite, hence the foliation in the quartz monzonite cannot be secondary.

The structure of this pluton of Kinsman quartz monzonite indicates that it is a thick intrusive sheet that moved into its present position while molten or partly molten. Its structure is analogous to that of plutons of Bethlehem gneiss in the Bellows Falls (Kruger,

1946, p. 190), Mascoma, Mt. Cube, and Moosilauke quadrangles (Billings, 1945, p. 59), which, like the body of Kinsman quartz monzonite in the Keene-Brattleboro area, lie to the west of plutons of the Oliverian magma series. It is believed that these bodies of Bethlehem gneiss, isolated by erosion, are western extensions of the Mt. Clough pluton, which lies to the east of the plutons of the Oliverian magma series. If the Kinsman quartz monzonite in the Winchester basin is a thick easterly dipping sheet, the roots of the sheet may well be one of the numerous steeply dipping bodies of Kinsman in the adjoining Monadnock quadrangle.

Brattleboro Syncline and Associated Minor Structures

The Brattleboro syncline, which trends about N. 10° E., occupies the area west of the Northey Hill thrust. Only the western limb of the syncline is present, for the eastern limb has been cut out by the Northey Hill thrust. The rocks west of the thrust appear to form an unbroken stratigraphic sequence, the top of which is toward the east. Most of the bedding and foliation dip steeply to the east, and the majority of the most reliable drag folds indicate that the top of the beds is toward the east.

Minor structural features in these rocks are much more complex than those in the metasediments around the domes of the Bronson Hill anticline. Minor folds of two distinct ages, each with an axial-plane foliation, are present in this part of the area.

Most of the older minor folds are isoclinal, or nearly so, and have an axial-plane flow cleavage. The flow cleavage cuts sharply across bedding on the noses of these folds, but on the limbs of the folds the foliation appears to parallel bedding. The axial planes of most of these folds strike northeast and are vertical or dip steeply east or west; a few of them are recumbent. The axes of most of these folds trend N. 3°–37° E., and most of them plunge NNE. or SSW. at angles less than 45° (areas 10, 13, and 16, Pl. 3). The attitude of many of these folds is modified by the younger folds.

The younger minor folds are more open and have an incipient to well-developed axial-plane slip cleavage. The cleavage becomes noticeable when platy minerals are bent into parallelism with the axial planes of these folds. More severe movement has resulted in closely spaced breaks parallel to the axial planes of the folds, and there has been some new mineral development parallel to these breaks. The axial planes of most of the younger minor folds strike slightly more easterly than do those of the older minor folds, and most of them dip 50°–80° NW. The axes of most of the younger minor folds plunge gently to steeply northeast or southwest. Minor folds of the younger generation cannot be used to indicate top and bottom of beds.

Minor folds of the two ages occur side by side in many outcrops, and in a few places the older folds have been refolded by the younger ones. At the corner of Union and Elliott streets, in Brattleboro, the axial plane of an older minor fold (Fig. 5) strikes N. 5° W. and has an average dip of about 40° E.; its axis plunges 11° N. 8° E. It is refolded by a younger fold whose axial plane strikes N. 90° E.; the dip of the axial plane and the plunge of the axis is 10° N. Other refolded folds are present in outcrops near the corner of Western Avenue and Cedar Street, Brattleboro.

Immediately north of the Brattleboro-Dummerston town line, along the highway north from Brattleboro, volcanics of the Orfordville formation strike about E.–W. and dip 25°–40° N. (Pl. 1). The rocks here are on the nose of a northerly plunging syncline, and beds would normally be younger toward the north. However, this fold is a large younger fold, and the beds are older toward the north. Younger minor folds here plunge N.–NE. at relatively low angles.

In many places the younger minor folds occur as closely spaced crinkles only a fraction of an inch in amplitude. In places the crinkles show an axial-plane slip cleavage. It is not unusual to find two, or even three sets of crinkles in one outcrop; the more prominent set, in general, has about the same orientation as larger younger minor folds in the same locality. The crinkles give a very pronounced lineation on bedding or flow cleavage.

In many instances it is impossible to decide whether a minor fold belongs to the younger or older group. The group of folds shown in the point diagrams as "unclassified folds" doubtlessly contains both younger and older folds.

The lower member of the Orfordville forma-

tion shows a characteristic type of minor folds. This member, composed of phyllite and interbedded thin quartzite and quartz-mica schist, contains many minor folds. In a single outcrop

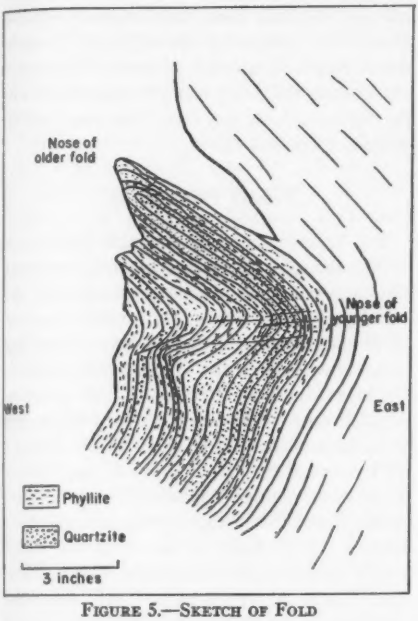


FIGURE 5.—SKETCH OF FOLD

one may find canoe-shaped doubly plunging folds, folds plunging north, and folds plunging south. In places, the pattern of adjoining folds indicates the top of the beds in opposite directions. Such folds are of little use in deciphering major structure.

FAULTS

General Statement

Numerous normal faults and one large thrust fault complicate the structure of the Keene-Brattleboro area. Most of the normal faults have a relatively small displacement; one, the Triassic border fault, has a maximum displacement of 15,000 to 20,000 feet. Three graben, in which metamorphic rocks are preserved, are present in the Westmoreland-Swanzey pluton. The Northey Hill thrust is an important structural break in western New Hampshire and eastern Vermont. Nearly all the faults trend parallel to the regional grain.

Northey Hill Thrust

The Northey Hill thrust has been mapped from the northern part of the Moosilauke quadrangle to the southwestern part of the Mascoma quadrangle (Billings, 1937; Hadley, 1942; Chapman, 1939). The thrust has not been mapped in detail in the Hanover and Claremont quadrangles but was recognized in the Bellows Falls quadrangle by Kruger (1946). The fault crosses the southeastern part of the Saxtons River quadrangle and enters the Keene-Brattleboro area just east of Bare Hill. From Bare Hill the thrust trends south and crosses to the eastern side of the Connecticut River about a mile upstream from Putney Station, where it swings south-southwest and recrosses the river about a mile upstream from Dummerston Station. The thrust then bends south, passes just east of Brattleboro, and extends beyond the southern edge of the area.

Postulation of a fault is necessitated by the great stratigraphic interval between the Orfordville formation and the Littleton formation. The position of the fault is locally revealed by a change in lithologic character or by structural discordance on opposite sides of the thrust, and by narrow bodies of highly sheared granitic rock which occur in a few places along the fault.

In most places along the fault in this area, argillaceous rocks with thin interbedded quartzites, belonging to the upper part of the Orfordville formation, have been brought into contact with argillaceous rocks of the Littleton formation, and it is very difficult to locate the fault accurately. The best structural evidence for the fault in the Keene-Brattleboro area is in the southeastern part of the township of Dummerston, where truncation of lithologic units is shown. Half a mile southwest of Dummerston Station, lime-silicate schist, amphibolite, and mica schist of the Littleton formation are in contact with the upper dark phyllite-quartzite member of the Orfordville formation. The rocks of the Littleton formation are on the eastern limb of a syncline and strike north or northeasterly. The fault at this locality strikes N. 30° E., about parallel to the foliation in the Orfordville formation. Just north of the Brattleboro-Dummerston town line, the upper member of the Orfordville formation has been cut out along the fault, and the slate member of

the Orfordville formation is in contact with schists of the Littleton formation.

The Northey Hill thrust, where it has been mapped farther north, is considered older than the climax of the regional metamorphism for in those areas there is no break in metamorphism, no fault breccia or obvious mylonitization, and no crushing of porphyroblasts associated with the fault (Billings, 1937; Chapman, 1939; Hadley, 1942; Kruger, 1946).

In the Keene-Brattleboro area the thrust has no topographic expression, no brecciation or fault gouge, and no silicification. However, mineral isograds are telescoped along the fault, and there has been a good deal of retrograde metamorphism near the fault. The garnet and the biotite isograds are very close to the fault except for a short distance at the northern edge of the area (Pl. 4). The telescoping of the isograds is best explained by post-metamorphism movement along the fault. In some places along, or near, the fault, garnet has been completely replaced by aggregates of chlorite which retain the crystal shape of the garnet. A fault along which there was post-metamorphism movement would offer relatively free passage to the water necessary for the alteration of garnet to chlorite; whereas a pre-metamorphism fracture would be healed during metamorphism and there should be no more retrograde action along it than there is elsewhere in the area. The intensely sheared bodies of granitic rock that occur along the fault suggest intermittent movement along the fault plane. These granitic bodies are probably members of the New Hampshire magma series, which was emplaced relatively late in the orogeny. The fault plane must have been a plane of weakness when the granitic rock was emplaced, and recurrent movement along the fault might well have caused the shearing in this rock. Hence, even though most of the movement along the fault occurred before the peak of metamorphism, there probably was considerable movement along the fault in this area after the climax of regional metamorphism.

Stratigraphic units known to be cut out by the fault in this area include an undetermined thickness of the upper part of the Orfordville formation, all of the Albee, Ammonoosuc, Partridge, and Clough formations, and a part of the Littleton formation. The stratigraphic throw is calculated to be at least 10,000 feet.

The fault surface was not observed in this area, but the relations between the fault trace and topography indicate that the fault dips steeply. The fault must have originally dipped toward the west, with the rocks on the west thrust relatively upward. If part of the movement along the fault was pre-metamorphism, the dip of the fault may have been considerably steepened by folding.

Triassic Border Fault

The Triassic border fault, which in this area strikes about N. 25° E. and extends from near the western foot of Hogback Mountain to the vicinity of Keene, where it apparently dies out, is the continuation of the fault that forms the eastern boundary of the Triassic rocks of northern Massachusetts. Although no fault gouge or silicified zone is exposed along the fault in this area, the presence of the fault is clearly revealed by truncated structure and by a sharp break in lithologic units on opposite sides of the fault. Along most of its extent in this area the fault follows the lowlands of the Ashuelot River, where outcrops are exceedingly scarce. However, Ammonoosuc volcanics, immediately east of the fault, are exposed 1.25 miles north of Hogback Mountain at an elevation of 640 to 660 feet in a small southerly flowing brook. The Ammonoosuc volcanics in this outcrop show a great deal of brown iron stain, are highly shattered, and show some alteration.

The fault cuts across the western and northern flanks of the Warwick dome, across the Stone Mountain syncline, and into the Westmoreland-Swanzey pluton. Kinsman quartz monzonite is present on the western side of the fault along most of its extent.

The western side of the fault is downthrown. If it is a high-angle normal fault in this area, as it is farther south, the fault plane must dip steeply northwest.

The maximum stratigraphic separation along the fault in the Keene-Brattleboro area is probably 15,000 to 20,000 feet. The thickness of the sheet of Kinsman quartz monzonite cannot be determined accurately, hence the figure for the stratigraphic separation is only approxi-

mate. The stratigraphic separation decreases to zero near the city of Keene.

The maximum stratigraphic separation along the Triassic border fault in the Keene-Brattleboro area is comparable to that of the fault farther south. The minimum stratigraphic separation along the eastern Triassic border fault in southern Connecticut is 16,000 feet, and the maximum may possibly be 30,000 feet (Russell, 1922).

The Triassic border fault is exposed at two places near French King bridge, 7 miles east of Greenfield, Massachusetts, about 10 miles south of the southern edge of the Keene-Brattleboro area (Keeler and Brainard, 1940). The fault in one exposure strikes N. 5° E. and dips 60° W.; a fault-line scarp extends N. 35° E. from the exposure of the fault plane. The fault in the second exposure strikes N. 48° E. and dips 35° NW.

Graben

General statement.—Three graben, in which metamorphic rocks are preserved, are present in the northwestern lobe of the Westmoreland-Swanzey pluton. One graben forms Hyland Hill, a second one is present in the Mine Ledge-Grays Hill area, and the third one forms West Hill. The metamorphic rocks in these graben are part of the roof of the Westmoreland-Swanzey pluton, and they show very clearly the structure of the roof of the pluton. Part of the roof of a pluton of Oliverian magma series, on top of a dome, is preserved in only two other localities in New Hampshire—on Smarts Mountain in the Mt. Cube quadrangle (Hadley, 1942), and in the northeastern part of the Mascoma quadrangle (Chapman, 1939).

The faults associated with these graben are very conspicuous. Not only is there structural discordance and changes in lithologic character on opposite sides of the faults, but in places the faults are expressed by silicified zones, brecciation, and fault-line scarps. The silicified zones along the faults are discontinuous, but in places they are striking features. Although they are narrow in most places, these zones have a maximum width of several hundred feet. Many of the silicified zones have topographic expression.

The upper surface of silicified zones, where preserved, are smooth planes and undoubtedly parallel the faults along which they occur. Numerous shear planes are present in some of the silicified zones. In general, they parallel the strike of the silicified zone, and most of them appear to dip in the same direction as does the silicified zone. Most of the shear planes in any one zone are undoubtedly roughly parallel to the fault along which they occur.

Some of the faults probably extend farther than indicated on the map. It is difficult to trace a fault where rocks of the Oliverian magma series are present on both sides of the fault, unless it is marked by a silicified zone.

The rock in the silicified zones is a hard white dull cherty quartz. Jasper occurs with the quartz in some places, hematite is occasionally present; and veinlets and vugs of drusy quartz are widespread. The silicified zone is the result of replacement of the country rock along the fault. All stages from unaltered to completely replaced rock may be seen in different places along the faults.

All the observed faults of these graben are normal faults. The dips range from 40° to 90°.

Mine Ledge graben.—The Mine Ledge graben, about 4 miles northwest of the city of Keene, is about 3 miles long and half a mile wide. The best exposure of one of the graben faults is in a cut along the Boston and Maine Railroad, about 900 feet west of the Keene-Surry town line, where the western boundary fault of the Mine Ledge graben is exposed. Quartz monzonite of the Oliverian magma series on the west side of the fault and amphibolite of the Ammonoosuc volcanics on the east side dip about 15° W. The fault plane at this locality strikes N. 37° E. and dips 55° SE. Fault gouge with a maximum thickness of about 25 feet is present along the fault, and the rocks on either side of the gouge show some brecciation, silicification, and chloritization. Patches and veins of vuggy quartz and calcite are present in the rocks for several hundred feet to either side of the fault. About 230 feet of Ammonoosuc volcanics is exposed on the east side of the fault; therefore the stratigraphic separation at this locality is at least 230 feet and is probably considerably greater. About a mile to the north,

the stratigraphic separation along the fault is calculated to be about 3000 feet.

The smooth upper surface of a silicified zone along the central fault of the Mine Ledge graben is exposed on the small 1200-foot knob, 2600 feet N. 60° W. of the site of the West Hill School. This surface strikes N. and dips 55° E.; several drusy quartz veins are about parallel to the face. Another silicified zone on the same fault, 1400 feet to the south, strikes N. 15° E. and dips 40° E. These two silicified zones are too small to show on the scale of the map. Ammonoosuc volcanics are present on the western side of the fault; rocks of the Oliverian magma series and the Ammonoosuc, Partridge, Clough, and Littleton formations are exposed along the eastern side of the fault. The maximum stratigraphic displacement along this fault is about 500 feet.

The eastern border fault of the Mine Ledge graben is the largest fault of this graben. A silicified zone several hundred feet wide and over a mile long forms Mine Ledge; smaller silicified zones occur at intervals along the fault. Jasper is an important constituent of the silicified zone at Mine Ledge. The rock also contains numerous veinlets of hematite, and iron ore is reported to have been mined from this locality. Rocks of various formations are exposed along the western side of the fault; granodiorite of the Oliverian magma series is exposed along the eastern side. The silicified zone on the 1100-foot knob just north of Dickinson Brook strikes N. 15°–30° E. and dips 70°–90° W. The maximum stratigraphic separation along this fault is 3000 to 4000 feet. The fault continues northward and extends at least 8 miles into the Bellows Falls quadrangle (Kruger, 1946).

Hyland Hill graben.—The Hyland Hill graben, about 6 miles west of the city of Keene, is very similar to the Mine Ledge graben. Silicified zones along the western fault indicate that the fault strikes about N. 10° E. and dips 60°–75° E. The maximum stratigraphic separation of this fault is about 3000 feet. The central fault is silicified, and some of the silicified gouge contains rounded fragments of quartz, schist, and volcanics. The fault, as indicated by the silicified zones, strikes N. 10°–45° E. and dips 60°–85° E. The maximum stratigraphic separation along the fault is about 1800 feet. The eastern border fault of this graben is somewhat sinuous. Silicified zones along the fault strike N. 8° W. to N. 22° E. and dip 70°–80° W. Fault breccia is conspicuous in some places along the fault. The maximum stratigraphic throw is about 1400 feet.

West Hill graben.—The West Hill graben, 1 miles southwest of the city of Keene, is complicated by numerous faults too small to show on the map (Pl. 1). The presence of many faults, eight of which are mapped, low dips, changing structure, and rough topography made it very difficult to predict outcrop patterns. This graben contains several faulted shallow synclines, each with a core of schist of the Littleton formation.

The eastern fault of this graben may be a branch of the Triassic border fault. Quartz diorite of the Westmoreland-Swanzey pluton is present on the eastern side of the fault; rocks of several formations are exposed on the western side. As indicated by silicified zones, the fault strikes N. 20° W. to N. 20° E. and dips steeply west. The maximum stratigraphic separation along the fault is probably several thousand feet.

The western side of this graben is not so clearly defined as the eastern side. The fault just west of Hill Brook is the most westerly fault whose downthrown side is to the east. This fault strikes N. 30°–75° E. and is not parallel to the general trend of the graben. Its average strike, between the faulted ends of the Clough quartzite on opposite sides of the fault, is N. 50° E. Numerous parallel joints along the fault near its southern end strike N. 30° E. and dip 60° SE. Numerous parallel shear planes in a silicified zone near the point where the fault crosses the Chesterfield-Swanzey town line strike N. 80° E. and dip 50°–65° S. By making several assumptions, the net slip of the fault can be calculated. Assuming that the dip of the fault is 60° SE. and that it is a dip-slip fault, the net slip of the fault is about 3100 feet.

The fault immediately west of Bailey Brook has a slightly sinuous trend which averages about N. 10° W. The smooth top of a silicified zone along the fault in one locality strikes N. 10° E. and dips 55° W. The western side of the fault is relatively downthrown; the maximum stratigraphic separation is about 1000 feet.

The other faults in this graben strike N. to N. 10° E., and along each the west side is relatively downthrown. The displacement along them varies, but none are larger than those just described. The large area mapped as silicified zone along the fault near the center of the graben (Pl. 1) is not as completely silicified as are some of the other areas shown as silicified zone. The rocks in this area show partial silicification, are crossed by a maze of vuggy quartz veinlets, and are highly shattered and brecciated.

Other High-angle Faults

Several high-angle normal faults are present around the flanks of the northwestern lobe of the Westmoreland-Swanzey dome; all show some silicification or brecciation. The displacement along these faults is probably relatively small. The north-south fault 2.5 miles east of the village of Spofford was recognized and followed in the granodiorite of the Oliverian magma series by the local silicified and brecciated zones along it. The displacement along the fault in the granodiorite may be considerably greater than it is farther south where the fault crosses metamorphic rocks.

The silicified zone near the northeastern corner of the area, in the body of Concord granite, is undoubtedly along a fault, but the fault could not be traced beyond the ends of the silicified zone, hence no fault is mapped.

Two parallel faults are present at the northern end of the Vernon dome. The downthrown side of each is to the west. The trace of the western fault follows a steep-sided gully. A steep face on the western side of the gully strikes N. and dips 65° E.; the face shows no slickensides and may be an erosional feature or a joint face. The steep face exposes breccia composed of angular to rounded fragments of schist, vein quartz, and quartzite in a matrix mainly quartz. The average fragment is 4–6 inches in diameter, but some are as much as 2 feet in diameter. Foliation in the slabs of schist shows random orientation. Fault breccia is also present in other places along the fault. The maximum stratigraphic throw along the fault is about 900± feet.

The age of the normal faults is discussed later.

METAMORPHISM

Progressive Metamorphism

The rocks of the Keene-Brattleboro area have been subjected to progressive dynamothermal metamorphism. The intensity of metamorphism increases both to the east and to the west of the general longitude of the city of Brattleboro (Pl. 4). Following the zonal classification of Harker, which has been used in recent studies of the metamorphic rocks of New Hampshire (Billings, 1935; Chapman, 1939; Hadley, 1942), the rocks in this area are divided into three zones, low-grade, middle-grade, and high-grade, based on the present mineral composition of the aluminous metasediments. The index mineral for the low-grade zone is chlorite; those of the middle-grade zone are biotite, garnet, and staurolite, in increasing order of metamorphic intensity; that for the high-grade zone is sillimanite. Surfaces which separate rocks of different metamorphic-intensity zones are mineral isograds. The staurolite isograd, for example, separates the rocks in which staurolite is present from those of a less intense zone in which staurolite is absent. In addition to the metamorphic intensity zones, Plate 4 shows isograds, which appear as lines on a map, for the different index minerals.

The rocks of the low-grade zone have been severely folded and sheared, whereas most of the rocks of the middle-grade and high-grade zones show relatively gentle folding. It is apparent that dynamic metamorphism alone was not responsible for the change in these rocks. The sillimanite isograd, which marks the boundary of the high-grade zone, curves several miles to the west, in the central part of the area, and follows very closely the western contact of the Kinsman quartz monzonite. This close association illustrates the important role of the New Hamsphire magma series in furnishing heat and solutions during the metamorphism of these rocks. The increase in intensity of metamorphism toward the western edge of the area may well be related to intrusives which are present only a short distance farther west.

Retrograde Metamorphism

Small amounts of retrograde minerals are present in many places in the Keene-Brattle-

boro area; however, most of the retrograde action has not been severe enough to change the rock materially. Chlorite, replacing minerals of higher-intensity zones, is the most common retrograde mineral. Sericite is present in small amounts in many of the rocks as an alteration product of plagioclase, garnet, staurolite, and sillimanite. Traces of calcite, which involve the introduction of CO_2, are present in some of the rocks.

More severe retrograde metamorphism is present along or near the Northey Hill thrust, where, in several places (Pl. 4), schists of the middle-grade zone have been highly chloritized and now resemble low-grade phyllite. Biotite and garnet porphyroblasts of these altered schists have been pseudomorphically replaced by chlorite or by an intergrowth of chlorite and sericite. Because the retrograde minerals contain more water than the replaced minerals, water must have had free access to rocks along the thrust during closing stages of metamorphism.

Occurrence of Minerals

The mineralogy of the metamorphic rocks of western New Hampshire and eastern Vermont has been discussed in more or less detail by others (Billings, 1937; Chapman, 1939; Kruger, 1946; White, 1946). The following paragraphs on the occurrence of minerals deals mainly with new data or with minerals that are not widespread in this metamorphic terrane.

QUARTZ: Quartz has a long range of stability, for it is found in all the metamorphic zones. All the rocks of the Keene-Brattleboro area except some of the lime-silicate rocks and some of the amphibolites contain quartz.

Kruger (1946, p. 196) states that in the Bellows Falls quadrangle:

"There is a conspicuous relationship between the abundance of quartz and the kind of porphyroblasts in the rock. Garnet occurs in rocks of all proportions of quartz and mica. The range is from 10 to 98 per cent quartz. Staurolite porphyroblasts are present in rocks containing 2 to 60 per cent quartz. The high-grade rocks containing sillimanite have from 55 to 90 per cent quartz."

As shown by Table 13, available data from the Keene-Brattleboro area indicate little or no difference in the range of the quartz content of rocks containing garnet, staurolite, or sillimanite porphyroblasts; further study might indicate a wider range in the quartz content of such rocks.

PLAGIOCLASE FELDSPAR: Table 14 gives the range in composition, and the average composition, of plagioclase from rocks of different metamorphic

TABLE 13.—AMOUNT OF QUARTZ IN ROCKS WITH DIFFERENT PORPHYROBLASTS

Kind of porphyroblast	Range of quartz content (Per cent)
Garnet	11–86
Staurolite	12–84
Sillimanite	2–90

zones in the Keene-Brattleboro area. Albite, and oligoclase as calcic as An_{18}, present in felsitic volcanics of the Orfordville and Waits River formations, are stable in the low-grade zone. Sericite, epidote, and traces of calcite were formed during the breakdown of the original plagioclase of the volcanics. Albite and oligoclase persist in the middle-grade zone if no anorthite or potential anorthite is present, and it presumably would be stable under such conditions even in the high-grade zone. The plagioclase of the middle-grade zone may contain any amount of anorthite, depending on the original composition of the rock. If the original rock is high in lime, as in the lime-silicate rocks, the resulting plagioclase is high in lime. There is little difference in the range of the anorthite content of the plagioclase in the schist and quartzite of the middle-grade and high-grade zones. Available data, however, show a tendency toward an increase in the average anorthite content of the plagioclase in the schist and quartzite of the high-grade zone over corresponding rocks of the middle-grade zone.

BIOTITE: Biotite is stable in potash-bearing rocks of the middle-grade and high-grade zones. It occurs as small well-oriented flakes and as porphyroblasts. Cross-biotite porphyroblasts are present in many of the rocks but are particularly conspicuous in parts of the Partridge formation, in the middle-grade zone, and in the Spaulding quartz diorite. These cross-biotites must have developed after shearing stresses ceased. The biotite associated with lime-silicate rocks of this area is purplish brown, notably different in color from the nearly black or reddish-brown biotite associated with the other rocks. The biotite of the lime-silicate rocks is high in magnesium and low in iron compared to the biotite in the more argillaceous rocks.

SILLIMANITE: Sillimanite is the index mineral of

METAMORPHIC ZONES OF THE KEENE-BRATTLEBORO AREA, NEW HAMPSHI

KEENE-BRATTLEBORO AREA, NEW HAMPSHIRE-VERMONT

Moore, Pl. 4

LEGEND

LOW-GRADE ZONE

MIDDLE-GRADE ZONE

HIGH-GRADE ZONE

AREAS OF RETROGRADE METAMORPHISM

Biotite

APPROXIMATE ISOGRADS

FAULTS

New Hampshire Magma Series

- cg – Concord granite
- sqd – Spaulding quartz diorite
- kqm – Kinsman quartz monzonite
- gd – Granodiorite and quartz diorite

Oliverian Magma Series

- olqm – Quartz monzonite
- olgd – Granodiorite
- olqd – Quartz diorite

- Dl – Littleton formation
- Sc – Clough quartzite
- Op – Partridge formation
- Oam – Ammonoosuc volcanics
- Oo – Orfordville formation
- Ow – Waits River formation

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the high-grade zone and is present in many of the stratified rocks in the eastern part of the area. It is also locally conspicuous in the Kinsman quartz monzonite; the excess alumina was presumably derived from inclusions of schist. Sillimanite is not confined to highly argillaceous rocks, for a thin bed of quartzite in the Ammonoosuc volcanics, on the northwest side of the 1300 foot knob about 1300 feet west of BM 1295 on the Pudding Hill Road, in the southeastern part of the area, contains about 8 per cent sillimanite as tiny needles in quartz grains and as bundles of fibrolite up to 4 mm. long and 2 mm. wide. The rock evidently did not contain enough soda, potash, or lime to use all the available alumina to form feldspar or muscovite. Sillimanite is also present in a quartzite pebble from the Ammonoosuc volcanics half a mile north of the above locality. Sillimanite replaces micas, feldspar, and quartz. Muscovite, in most cases, appears to be the first mineral replaced. Some sillimanite has been replaced by muscovite during retrograde metamorphism.

Table 14.—Composition of Plagioclase in Different Metamorphic Zones

Zone	Type of rock	Number of determinations	Anorthite content of plagioclase	
			Range	Average
Chlorite (low-grade)	Felsitic volcanics	4	10–25*	15
Garnet (middle-grade)	Light volcanics (Biotite gneiss)	12	9–29	24
	Dark volcanics (Amphibolite)	4	20–40	30
	Schist and quartzite	5	6–26	15
	Lime silicate	3	25–95	68
Staurolite (middle-grade)	Light volcanics (Biotite gneiss)	1	19	19
	Schist and quartzite	7	13–29	25
Sillimanite (high-grade)	Light volcanics (Biotite gneiss)	1	26	26
	Schist and quartzite	2	26–32	29

* An_{95} is probably a relic mineral.

AMPHIBOLE: Two types of amphibole are common in the metamorphic rocks of this area (Table 15). One type is light green to gray green in hand specimen; X = colorless to yellow, Y = light green to colorless, Z = medium green to colorless, in thin section; β = 1.641 to 1.657, $Z \wedge c$ = 20°–24°. This amphibole, present in lime-silicate beds of the Littleton and Partridge formations in the middle-grade zone, and in some beds of light-colored biotite gneiss of the Ammonoosuc volcanics in the middle-grade and high-grade zones, may be actinolite in spite of its high extinction angle. If it is actinolite, Winchell's tables (1933, p. 246) indicate that it is intermediate between tremolite and ferrotremolite. The actinolite in the lime-silicate beds is associated with quartz, intermediate to calcic plagioclase, zoisite, diopside, and garnet. The lime-silicate beds were originally dolomitic shale or sandstone. The actinolite was formed by a reaction between dolomite and quartz (Billings, 1937, p. 546). As neither calcite nor dolomite was observed in the lime-silicate beds, equilibrium conditions must have been reached in these rocks.

Sufficient data are not available to map an actinolite isograd. It is widespread in the lime-silicate rocks of the garnet zone and it may form under less intense conditions of metamorphism than does garnet.

Interbedded with the light-colored biotite gneiss of the Ammonoosuc volcanics are black amphibolites. The amphibole in the amphibolite is common hornblende; it shows X = yellowish-green, Y = green, and Z = blue-green in thin section; β = 1.674 to 1.681, and $Z \wedge c$ = about 20°. The distribution of the two types of amphibole in the Ammonoosuc volcanics is undoubtedly due to differences in chemical composition of the original volcanics.

The amphibole in the plutonic rocks is common hornblende with optical properties similar to that in the amphibolite.

ANTHOPHYLLITE: Anthophyllite was recognized in

TABLE 15.—OPTICAL PROPERTIES OF AMPHIBOLES OF DIFFERENT FORMATIONS

Formation	Color	Pleochroism	Index			Z ∧ c
			α	β	γ	
Ammonoosuc volcanics, biotite gneiss	Light green	None		1.651		20°
Ammonoosuc volcanics, biotite gneiss	Very light green	None	1.642	1.657	1.671	20°
Ammonoosuc volcanics, amphibolite	Black	X-straw yellow Y-grass green Z-blue green	1.670	1.681	1.688	20°
Ammonoosuc volcanics, amphibolite	Black	X-straw yellow Y-grass green Z-blue green	1.662	1.674	1.681	20°
Ammonoosuc volcanics, amphibolite	Black	X-yellow green Y-dark green Z-blue green		1.676		19°
Partridge formation, lime-silicate bed	Dark gray green	X-straw yellow Y-light green Z-light blue green	1.638	1.651	1.662	21°
Littleton formation lime-silicate beds	Light green	X-colorless Y-light green Z-medium green	1.637	1.651	1.662	24°
Littleton formation lime-silicate bed	Gray	X-colorless Y-colorless Z-very light green	1.629	1.641	1.648	24°
Littleton formation lime-silicate bed	Gray-green	X-colorless Y-very light green Z-light green	1.627	1.642	1.651	23°
New Hampshire magma series amphibolite sill	Black	X-light green Y-dark green Z-dark blue green	1.663	1.676	1.683	29°
New Hampshire magma series amphibolite sill	Black	X-yellow green Y-dark green Z-dark blue green	1.676	1.686	1.692	27°
Oliverian magma series Vernon pluton	Black	X-yellow green Y-dark green Z-blue green	1.668	1.677	1.685	22°

several localities in the Ammonoosuc volcanics of the high-grade zone. At one locality, in the southeastern part of the township of Winchester, on the east side of the center knob 0.65 mile south of Wheeler Mountain, coarse anthophyllite schist is present in a green to black biotite-rich schist interbedded with light-colored biotite gneiss. The anthophyllite, which occurs as lamellar prisms up to 4 cm. long, has a γ index of 1.678. This, according to Rabbitt (1946, Table facing p. 41), indicates either

a $FeO + Fe_2O_3 + TiO_2 + MnO$ content of about 25 per cent or possibly a rather high alumina content. Small grains of blue kyanite are scattered throughout the anthophyllite, suggesting an original sediment high in alumina. Emerson (1917, p. 74) gives an analysis of gedrite, a variety of anthophyllite, from the Dana diorite (Ammonoosuc volcanics) around the Monson granodiorite (Oliverian magma series), from Richmond, Vermont (*sic*) which contains 14.51 per cent $FeO + Fe_2O_3 + TiO_2 + MnO$ and 14.09 per cent Al_2O_3. The complete analysis, as given, is:

	1	2
SiO_2	47.86	50.65
TiO_2	.63	.50
Al_2O_3	14.09	13.03
Cr_2O_3	—	tr
Fe_2O_3	.33	.27
FeO	13.41	12.67
MnO	.14	.15
CaO	.57	1.73
MgO	19.89	16.96
K_2O	.06	.04
Na_2O	.93	1.37
H_2O	2.46	2.96
P_2O_5	.05	tr
Total	100.42	100.33

1. "Gedrite, fresh material. From Dana diorite, eruptive." (Presumably fresh uncontaminated material)

2. "Gedrite, average rock." (Presumably unselected material)

The gedrite Emerson analyzed is probably from Richmond, New Hampshire, where the geology is the same as that described for the gedrite locality, instead of from Richmond, Vermont, which is in northwestern Vermont. The township of Richmond, New Hampshire, is less than 1 mile east of the anthophyllite locality described in this paper.

KYANITE: In addition to the occurrence with anthophyllite in Winchester, kyanite occurs in several places in the area as a hydrothermal mineral. Blades of kyanite, as much as 15 mm. long, are concentrated along or near pods of quartz in biotite gneiss of the Ammonoosuc volcanics on the northwestern slope of Bald Hill, in Westmoreland Township, at an elevation of about 1100 feet. The kyanite has been partially replaced by sericite during retrograde metamorphism.

Large blades of blue kyanite are present along the walls of a quartz vein in amphibolite of the Ammonoosuc volcanics on the south side of the 1240-foot knob just south of the Westermoreland-Chesterfield town line on the Poocham Road, in the northeastern part of the area. The alumina may have been introduced or may have been derived from the wall rock; regardless of the source of the alumina, the kyanite in this occurrence must be a hydrothermal mineral.

Sufficient data are not available to map a kyanite isograd.

CORDIERITE: A very coarse amphibolite, highly altered to talc and chlorite, (Table 3, column 12) is present in the Ammonoosuc volcanics along the western side of Sprague Brook, in the southern part of the town of Richmond. Massive dark-blue cordierite is exposed along the edges of a quartz vein which cuts the altered amphibolite at a locality 1.7 miles S.40°W. from the village of Richmond. The optical properties of the cordierite are as follows: $(-)\ 2v = 75\pm$; $\alpha = 1.545$, $\beta = 1.549$, $\gamma = 1.552$. The cordierite in this locality appears to be hydrothermal, formed as a replacement of the country rock along the quartz vein. Hitchcock (1878, vol. 3, part 4, p. 77) gives the following analysis of cordierite (iolite) from the town of Richmond, New Hampshire:

	(Per cent)
SiO_2	48.0
Al_2O_3	35.0
FeO	6.0
MgO	10.0
MnO	1.0
Total	100.0

TRIASSIC (?) MINERALIZATION

The age of the fluorite-bearing veins in the towns of Westmoreland and Chesterfield is of interest because of the evidence it offers concerning the age of other structural features in the area. The location of mines on two of the larger veins, 1.5 miles S. 25° E. of Poocham near the Westmoreland-Chesterfield town line, is shown on Plate 1. Bannerman (1941) has described the fluorite deposits in detail.

The fluorite veins occur in gneiss of the Oliverian magma series and in the overlying Ammonoosuc volcanics; all the veins are close to the contact between these two rocks. The vein material was deposited as fissure fillings in steeply dipping fractures. The strikes of the observed veins are N. 22° E. to N. 45° E. The average strike of the group of veins is about

N. 30° E., and all but one dip SE. The veins parallel a well-developed set of joints in the country rock. In some places, narrow altered or silicified zones are present in the wall rock along the veins. The minerals present, in order of their abundance, are quartz, fluorite, barite, and calcite, with minor amounts of pyrite, chalcopyrite, galena, sphalerite, and bornite.

Fluorite-bearing veins cut Triassic and older rocks in a wide band along the Connecticut River between Northampton and Greenfield, Massachusetts. These veins are similar in many respects to the fluorite deposits of Westmoreland and Chesterfield (Emerson, 1895, p. 82; 1898, p. 444, 502).

Because the Massachusetts deposits resemble so closely the Westmoreland-Chesterfield fluorite veins, the author believes the latter are also Triassic or younger.

Age of the Graben and Normal Faults

It appears to be possible to date the graben and normal faults of the Keene-Brattleboro area by certain similarities with high-angle normal faults in the Triassic rocks of the Connecticut Valley.

Pertinent data concerning the faults in the Keene-Brattleboro area are as follows:

(1) The eastern Triassic border fault trends about N. 25° E. from neas Hogback Mountain to the vicinity of Keene, where it may die out or may swing to the north and continue as the easternmost fault of the West Hill and Mine Ledge graben.

(2) Most of the faults of the graben strike N. to N. 10° E. and have relatively steep dips.

(3) Most of the normal faults around the flanks of the domes strike about N. 30° E.; the two at the northern end of the Vernon dome strike N. 10° W. Available data indicate that they all dip steeply.

(4) The graben faults and most of the other normal faults are strongly silicified or brecciated in places.

(5) These high-angle normal faults are west of the Triassic border fault.

(6) The close parallelism between the normal faults and graben, the fractures along which the Westmoreland-Chesterfield fluorite veins occur, and the strong jointing in the rocks around the fluorite deposits suggest a common origin and common age for these features. It has been pointed out that the fluorite veins are probably Triassic or younger.

The Triassic rocks of the Connecticut Valley occur in a large tilted fault block. Longwell (1937) has shown that, in Connecticut, the eastern border fault was active during Triassic sedimentation. Movement along the entire fault is probably of the same age, and therefore the movement along this fault in the Keene-Brattleboro area occurred during the Triassic.

Near French King bridge, 7 miles east of Greenfield, Massachusetts, the eastern Triassic border fault is characterized in places by slickensides and crushed and brecciated zones impregnated with hematite (Keeler and Brainard, 1940, p. 362). These same features are characteristic of most of the normal faults in the Keene-Brattleboro area. Hematite is particularly abundant along the eastern fault of the Mine Ledge graben at Mine Ledge, about 5 miles northwest of the city of Keene.

Along most of its extent in Connecticut and Massachusetts, the eastern Triassic border fault trends N. to N. 20° E. and dips steeply west. The Triassic rocks, west of the border fault, are cut by many small normal faults. Bain (1941, p. 270) presents evidence that movement along some of these smaller faults occurred during Triassic sedimentation. In central and northern Massachusetts most of these faults trend N. to N. 30°E. and have steep dips; some of them bound graben (Bain, 1941, map facing p. 264; Wheeler, 1937, Pl. 1). The map pattern of these faults in the Triassic rocks is like the pattern of the smaller normal faults and graben in the Keene-Brattleboro area. It seems logical that the same type of faulting, bearing similar relations to the same master fault, and in localities only a short distance apart, is of the same age. The border fault does not stop at the most northerly exposure of Triassic rocks, and there is no obvious reason why the small normal faults should do so. For these reasons, the normal faults and graben of the Keene-Brattleboro area are believed to be Triassic.

This conclusion further indicates that the normal faults that trend NNE. through western New Hampshire at least as far north as the Mt. Washington quadrangle are also Triassic.

FIGURE 1. WAITS RIVER FORMATION
Smooth dark bed on right is quartz-muscovite marble, beds on left are biotite schist. 2.5 miles N. 20° E. of the city of Brattleboro.

FIGURE 2. INTERBEDDED QUARTZITE AND SLATE OF THE ORFORDVILLE FORMATION
In the schoolyard, city of Brattleboro. Bedding strikes N., dips 75° E. A slip cleavage approximately parallel to the axial planes of the shallow folds strikes N. 27° E. and dips 75° NW.

FIGURE 3. GRANODIORITE OF OLIVERIAN MAGMA SERIES WITH LARGE TABULAR INCLUSION OF AMMONOOSUC VOLCANICS
At the dam on the Ashuelot River at northern edge of the area.

FIGURE 4. GRANODIORITE OF OLIVERIAN MAGMA SERIES WITH LARGE INCLUSIONS OF AMMONOOSUC VOLCANICS
Both cut by light-colored pegmatite dikes. At the dam on the Ashuelot River at the northern edge of the area.

OUTCROPS

FIGURE 1. PHYLLITE OF ORFORDVILLE FORMATION
Flow cleavage dips toward the reader, slip cleavage dips to right. Incipient slip cleavage expressed as crinkles that produce a prominent lineation.

FIGURE 2. CRINKLED SCHIST OF THE LITTLETON FORMATION
With interbedded thin quartzites. Flow cleavage and bedding are parallel.

FIGURE 3. LIGHT-COLORED BIOTITE-QUARTZ-FELDSPAR GNEISS OF AMMONOOSUC VOLCANICS

FIGURE 4. GRANODIORITE OF WESTMORELAND-SWANZEY PLUTON
Shows prominent biotite clusters.

FIGURE 1. PORPHYRITIC AMPHIBOLITE CONTAINING GRANULATED AND DRAWN-OUT PHENOCRYSTS OF WHITE PLAGIOCLASE

FIGURE 2. QUARTZ DIORITE FROM SMALL STOCK 0.5 MILE NORTHWEST OF VILLAGE OF WESTMORELAND

FIGURE 3. PORPHYRITIC KINSMAN QUARTZ MONZONITE

FIGURE 4. SPAULDING QUARTZ DIORITE

PLUTONIC ROCKS OF THE NEW HAMPSHIRE MAGMA SERIES

FIGURE 1. PORPHYRITIC KINSMAN QUARTZ MONZONITE
Shows inclusions of schist of Littleton formation. Both the inclusions and the feldspar phenocrysts parallel foliation in the quartz monzonite. At road corner marked X700, 1.4 miles southeast of Fullam Pond. Photo by R. F. Meyer.

FIGURE 2. QUARTZ DIORITE OF OLIVERIAN MAGMA SERIES
Containing reworked inclusions of Ammonoosuc volcanics. Inclusions parallel foliation of the plutonic rock. 0.25 mile north of village of Richmond.

FIGURE 3. FOLDED CONTACT BETWEEN AMMONOOSUC VOLCANICS (dark band near base and center of picture) AND GRANODIORITE GNEISS OF OLIVERIAN MAGMA SERIES (light rock at right)
Foliation in gneiss parallels axial plane of fold but cuts across bedding of volcanics. Axial plane of fold strikes N. 90° E., dips 25°–30° N. In road cut 0.5 mile north of Hogback Mountain (3 miles S. 25° W. of village of Winchester).

FIGURE 4. FOLD IN AMMONOOSUC VOLCANICS
Bedding and foliation, mutually parallel, wrap around nose of fold. Thin white band is aplite. Flattened pebbles can be seen in some of the beds. In road cut 75 feet west of contact shown in Figure 3.

OUTCROPS

Structural Evolution

Sedimentation dominated from middle Ordovician (?) through early Devonian. About 4000 feet of marine mud, sand, and impure calcareous ooze, with some felsitic volcanics, accumulated to form that part of the middle Ordovician (?) Waits River formation which is present in this area. The accumulation of the subsequent 4000± feet of mud and sand, which form the Orfordville formation, was accompanied by extensive volcanic activity.

During the late Ordovician (?), sediments of the Albee formation were presumably deposited in this area. In the Mt. Cube area (Hadley, 1942), the Albee consisted of 5000± feet of marine sand, mud, and a little quartzose gravel. Subsequently, extensive volcanic activity resulted in the accumulation of more than 1000 feet of alternating felsitic and basaltic tuff, breccia, and flows which form the Ammonoosuc volcanics. A minimum of 1000 feet of mud and sand, which constitutes the Partridge formation, was deposited above the volcanics. The deposition of the mud and sand was accompanied by short periods of mild vulcanism.

A period of mild orogeny and erosion interrupted sedimentation during late Ordovician (?). In parts of the area the Partridge formation is absent, probably as a result of the late Ordovician (?) erosion rather than because of a lack of original sedimentation.

Sand and quartz gravel, which form the Clough quartzite, accumulated in the early Silurian sea. In places these sediments were at least 1000 feet thick, however, over most of the area they were either thin or were largely removed by erosion during middle or late Silurian.

The sediments deposited to form the lower Devonian Littleton formation of this area were highly variable. The majority of the sediments were mud or arenaceous mud with many thin beds of sand. Thin beds of quartz gravel were deposited over parts of the area. Highly calcareous or dolomitic sands, some associated with beds of quartz gravel, were deposited at several horizons. Flows and volcanic tuff are present at several horizons; the thickest are near the base of the formation. The total thickness of these sediments was at least 5000 feet.

Some time after the deposition of the Littleton formation, during early stages of the late Devonian orogeny, the rocks were intruded by the Oliverian magma series.

Billings (1937, p. 502) believed that the intrusion of the Owls Head granite (Oliverian magma series) domed the surrounding sediments. Chapman (1939, p. 155) considered the bodies of the Oliverian magma series to be laccoliths and believed that the surrounding concentric schistosity, and lineation, was produced by the upward push of the magma combined with the downward force due to load. In the Mascoma dome the elongation of the roof, owing to laccolithic intrusion, was calculated to have been at least 12 per cent of its original length (Chapman, 1939, p. 156).

Assuming that the ends of the rock section at the periphery of the domes remained stationary—that is, they did not move toward each other during the formation of the domes—the calculated elongation of the roof of the northwestern lobe of the Westmoreland-Swanzey dome, as shown by its geometry, is about 6 per cent; that of the Warwick dome is about 26 per cent. In testing Chapman's theory of laccolithic intrusion, the elongation shown by the roof of an overturned dome must be discounted, for Chapman considered such overturn to be due to later folding (Chapman, 1939, p. 157).

The percentage of distortion of elongate pebbles in the Clough quartzite around the domes of the Keene-Brattleboro area has been calculated, for representative pebbles from eight widely distributed areas, according to the method used by Cloos (1943, p. 279). Assuming that the pebbles were originally spheres, the percentage of extension in the direction of the long axis of the pebbles ranges from 65 to 280 per cent; the change in the intermediate axis ranges from a shortening of 19 per cent to a lengthening of 23 per cent; and the shortening of the least axis of the pebbles ranges from 45 to 69 per cent.

Comparison of the elongation shown by stretched pebbles in the Clough quartzite with the elongation of the roof of the domes shows that radial stretching of the beds due to laccolithic intrusion is by no means sufficient to produce the elongation of the pebbles. That is, whereas the actual lengthening of the pebbles

ranges from 65 to 280 per cent, the lengthening due to laccolithic intrusion would be only 6 to 26 per cent.

It is concluded, therefore, that neither the domes, the surrounding concentric schistosity, nor the associated lineation was produced by laccolithic injection of the Oliverian magma series.

Hadley (1942, p. 162, 163) concluded that the plutons of the Oliverian magma series are syntectonic phacoliths; that the foliated structures are due primarily to doming contemporaneous with the intrusion; that the linear structures in both the roof rocks and the intrusives, produced simultaneously with the foliation, are closely related to the regional deformation; and that recrystallization under stress essentially normal to the beds oriented mica flakes and kyanite rods, and helped to flatten and elongate quartz pebbles.

Chapman (1942, p. 907, 908) presents the following two objections to the phacolithic theory of injection:

> (1) "The arches of the domes are too broad and open to permit the formation of phacoliths much thicker than several hundred feet. It is obvious that the domes have been eroded so much that the floors of such phacoliths would have been exposed long ago and most of the igneous body destroyed."
> (2) "If the domes had developed under regional compression and the magma flowed in to form phacoliths, there would have been practically no shearing or slipping between and within beds in these broad open arches. The writer believes that some rock flowage, accomplished by slipping and shearing during the doming, was essential in the formation of the bedding plane schistosity."

It is concluded that phacolithic intrusion of the Oliverian magma series is not competent to explain the structure in and around these plutons.

The writer believes that the Westmoreland-Swanzey pluton, the Vernon pluton, and the Warwick pluton, all belonging to the Oliverian magma series, were forcefully injected as one thick composite sheet during very early stages of the orogeny, while the enclosing rocks were essentially flat. Proof that the intrusion is locally older than the folding is given in the description of the structure of the Warwick dome. Proof that this is true on a large scale is based on regional relations. The Oliverian magma series intrudes approximately the same horizon for at least 140 miles along the strike of the Bronson Hill anticline. The rocks above and immediately surrounding these plutons show very simple structure and low dips, whereas the rocks between the plutons are tightly and complexly folded. If the rocks were highly folded before the intrusion of the Oliverian magma series it would be impossible to explain these simple structural relations.

The foliation in the rocks of the Oliverian magma series, with local exceptions, must be a primary foliation because: (1) In the Keene-Brattleboro area, the rocks of the Oliverian magma series contain shear zones along which the foliation has been slightly displaced. Many of the shear zones are filled with pegmatitic quartz, microcline, biotite, and magnetite which grade into the country rock and which appear to be segregation pegmatites formed by the draining of the residual liquid into the shears. (2) Numerous elongate inclusions of Ammonoosuc volcanics are oriented parallel to the foliation in the enclosing rock. (3) The structure of the plutons is conformable with that of the overlying stratified rocks. (4) The rocks of the Oliverian magma series in the Westmoreland-Swanzey pluton had a directional weakness (foliation) before the amphibolite sills and dikes of the New Hampshire magma series, which are pre-metamorphism, were intruded. The rocks of the pluton, furthermore, were cool, for the sills and dikes show chilled borders.

The primary foliation in the rocks of the Oliverian magma series is believed to have formed as a result of forceful injection of a partly crystallized magma, resulting in flow layers that oriented the biotite flakes into the plane of the flow layers.

The flat-lying primary foliation, parallel to the bedding of the country rocks, responded to subsequent regional folding in the same manner as did the bedding-plane surfaces of the stratified rocks. Planes of primary foliation were planes of weakness during subsequent folding, and in most places shearing and slipping due to regional compression occurred along the planes of primary foliation and produced the characteristic granoblastic texture of the Oliverian magma series.

It seems clear that the foliation, and the

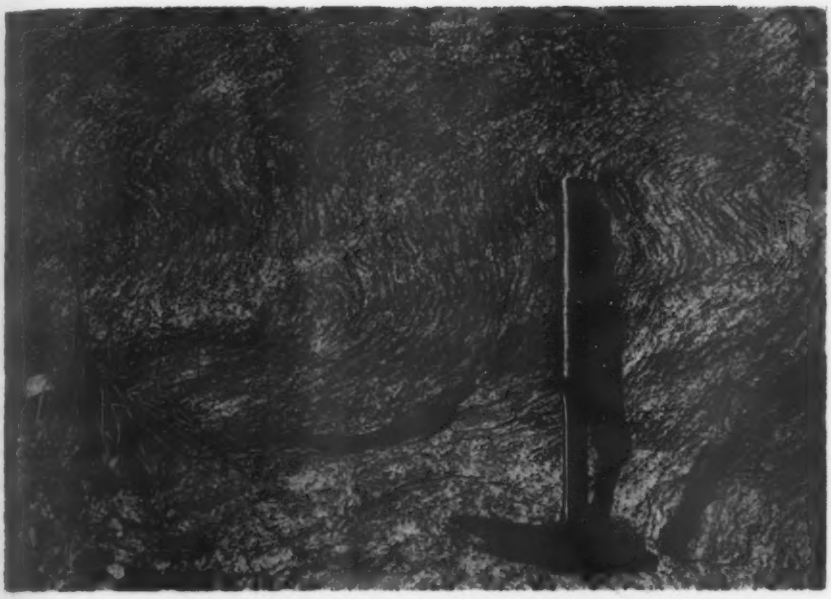

FIGURE 1. TYPICAL GNEISS OF OLIVERIAN MAGMA SERIES
Shows shear zones, along which foliation is slightly offset, filled with pegmatitic material. In road cut 0.5 mile north of Westport. Photo by James A. Haertlein.

FIGURE 2. LENS-SHAPED PEGMATITE IN ROCKS OF LITTLETON FORMATION
Near Mine Ledge in Surry. Photo by James A. Haertlein.

lineation expressed by stretched pebbles and elongate minerals, in the stratified rocks, formed simultaneously during regional folding. As the lineation of the Oliverian magma series is parallel, in most places, to the lineation in adjacent stratified rocks, it appears that they, also, developed simultaneously. A secondary foliation was locally developed, as at the locality described under the structure of the Warwick dome where the foliation in the Oliverian magma series is parallel to the axial plane of a fold shown by the contact of this rock with the Ammonoosuc volcanics.

During the late Devonian orogeny, subsequent to the intrusion of the Oliverian magma series, the Ordovician (?), Silurian, and Devonian stratified rocks were folded by compressive forces acting in an east-west direction. The Bronson Hill anticline and the Brattleboro syncline began to develop, and domes were buckled up on the Bronson Hill anticline. The foliation in the stratified rocks and the lineation in both the stratified rocks and in the plutons of the Oliverian magma series were formed at this time. The foliation in the stratified rocks above and immediately around these plutons developed as a bedding-plane foliation, and the igneous bodies acted as buttresses that prohibited tight folding in the overlying rocks. Elsewhere tight folds were formed and an axial plane flow cleavage developed. The folds and the cleavage of this age are the older folds and cleavage, as previously discussed. Initial movements along the Northey Hill thrust may have occurred at this time; but a considerable part of the movement along the thrust occurred after the metamorphic intensity in the rocks along the fault had reached its maximum, for in the Keene-Brattleboro area mineral isograds are telescoped along the fault. Progressive dynamothermal metamorphism was probably instigated at this time, but the peak of metamorphism was not reached until the Kinsman quartz monzonite of the New Hampshire magma series was intruded. The lineation of the pebbles in the Clough quartzite was produced by horizontal compression which squeezed them out of their original shape, a process discussed by Billings (1942, p. 235). The lineation of the pebbles, whose short axes are perpendicular to the bedding, shows that there has been thinning of the beds and extension of them parallel to bedding planes. The upward lengthening of the rock was accomplished by folding. Horizontal lengthening of the pebbles at right angles to compression was accomplished by the lengthening of individual beds in that direction. Lengthening of the beds parallel to the axes of the folds is demonstrated by salients and plunging folds, for both necessitate stretching parallel to the axes of the folds. Mineral lineation and the foliation in the roof rocks can be explained in the same way.

Porphyritic basaltic sills and dikes of the New Hampshire magma series were injected into the southern part of the Westmoreland-Swanzey pluton after the rocks of that body were relatively cool, but before the peak of regional deformation had been reached, for the basaltic sills and dikes have been recrystallized to form amphibolite. Stocks and sills of granodiorite and quartz diorite, the Kinsman quartz monzonite, the Spaulding quartz diorite, and the Concord granite, all of the New Hampshire magma series, were intruded in the order listed (Table 12). The granodiorite and quartz diorite were intruded while regional stresses were still active, for the quartz of these intrusive rocks is highly granulated, but after the metasediments were highly folded, for the granodiorite and quartz diorite contain folded inclusions of the country rock.

The Kinsman quartz monzonite was forcefully injected as a thick sheet after the stratified rocks had been highly folded. The Kinsman contains inclusions of folded country rock, and the schistosity of some of the inclusions is not parallel to the foliation in the Kinsman. Because the phenocrysts in the quartz monzonite show very little granulation, the Kinsman must have been intruded near the close of the period of folding. The foliation in the Kinsman quartz monzonite, shown by biotite and muscovite and to a lesser extent by the phenocrysts, is a primary foliation. The easterly dip of the Kinsman quartz monzonite and the structural similarity of this pluton with several bodies of Bethlehem gneiss farther north in New Hampshire suggest that the roots of this pluton lie to the east, possibly in the adjacent Monadnock quadrangle. Each successive intrusion of

rocks of the New Hampshire magma series is less deformed; the last member of the series in this area, the Concord granite, is not granulated and shows only weak foliation.

Although metamorphic minerals probably started to form very early in the orogenic period, the peak of metamorphic intensity was not reached until emplacement of the Kinsman quartz monzonite. Heat and solutions furnished by the Kinsman quartz monzonite were very important in the process of metamorphism in this area, for the zone of most intense metamorphism is immediately around, and above, the body of Kinsman quartz monzonite.

The younger folds and the attendant axial-plane slip cleavage formed during the decline of metamorphic intensity, for there has been very little recrystallization or formation of new minerals along the slip cleavage. Either compressive stresses acting in a northwest-southeast direction or a couple in which the rocks to the east moved northward may be responsible for the younger folds and cleavage.

Further movements along the Northey Hill fault recurred after the peak of metamorphic intensity. Possibly at the same time some of the metamorphic minerals suffered retrograde metamorphism. Some of the retrograde metamorphism appears to be due to a regional decline in intensity conditions, but the retrograde metamorphism along the Northey Hill thrust appears to be a local phenomenon due to hydrothermal solutions.

Post-orogenic igneous activity, which was widespread in parts of New Hampshire, was not important in the Keene-Brattleboro area. One post-metamorphism mafic dike was observed, but it is so weathered it was impossible to decide whether it belongs to the White Mountain magma series or is one of the Triassic basaltic dikes.

Triassic sediments crop out near the Connecticut river 1 mile south of the Keene-Brattleboro area. Probably these sediments formerly extended northward into the Keene-Brattleboro area. During Triassic time a large normal fault, the eastern Triassic border fault, which extends at least from the city of Keene southward to Long Island Sound, cut across the rocks of the eastern part of the area. The maximum stratigraphic throw along this fault in the Keene-Brattleboro area is 15,000 to 20,000 feet. Three graben and a number of small normal faults, probably Triassic, are present in the area. The abundance of Triassic block faults in southern New England indicates stretching in the associated rocks. Most geologists have concluded that such stretching is a result of tension, though this does not preclude the possibility that the tension is the indirect result of compressional, rotational, or torsional forces (Billings, 1942, p. 210). The writer believes that the most logical explanation of the normal faults and graben in the Keene-Brattleboro area is that they were formed by differential vertical movements which resulted in stretching the associated rocks. The rocks were brittle enough to fail by rupture. The fluorite deposits of the towns of Westmoreland and Chesterfield appear to be closely related to the faults and are also probably Triassic.

The Keene-Brattleboro area, in general, has been subjected to erosion since the close of the Devonian. Erosional processes may have been interrupted during part of Triassic time, and were temporarily interrupted during part of Pleistocene time. Pleistocene glaciers covered the area and deposited a thin layer of till. Temporary lakes covered that part of the area along the Connecticut River and the area immediately around the city of Keene (Hitchcock, 1878, p. 67).

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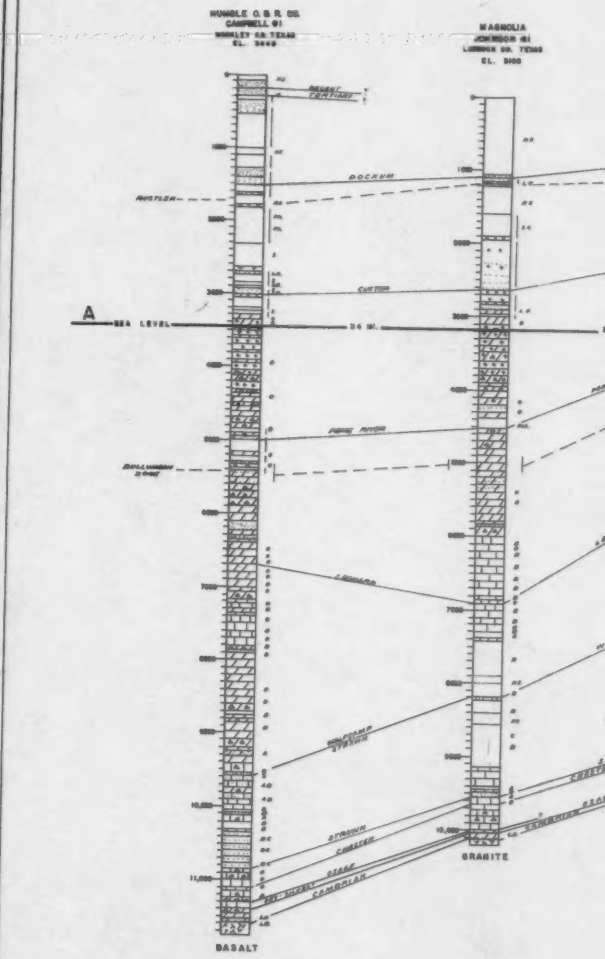
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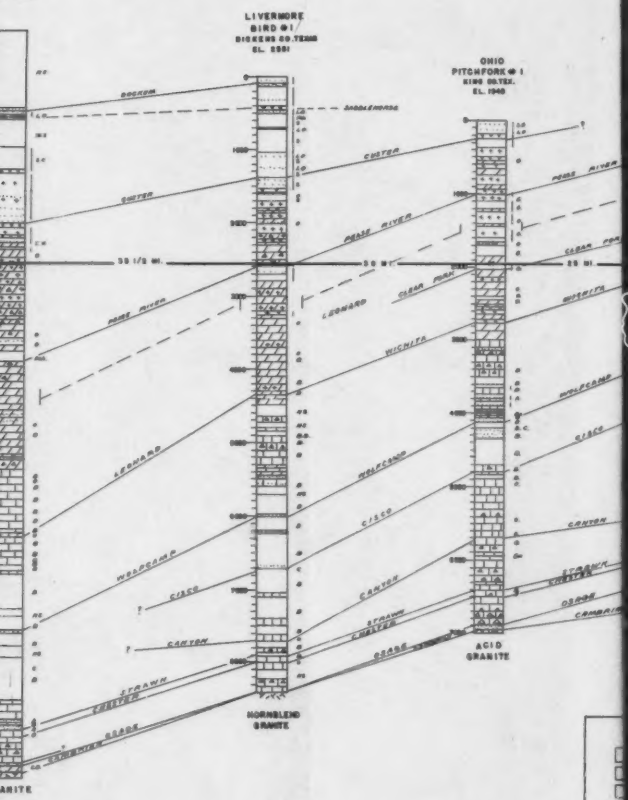
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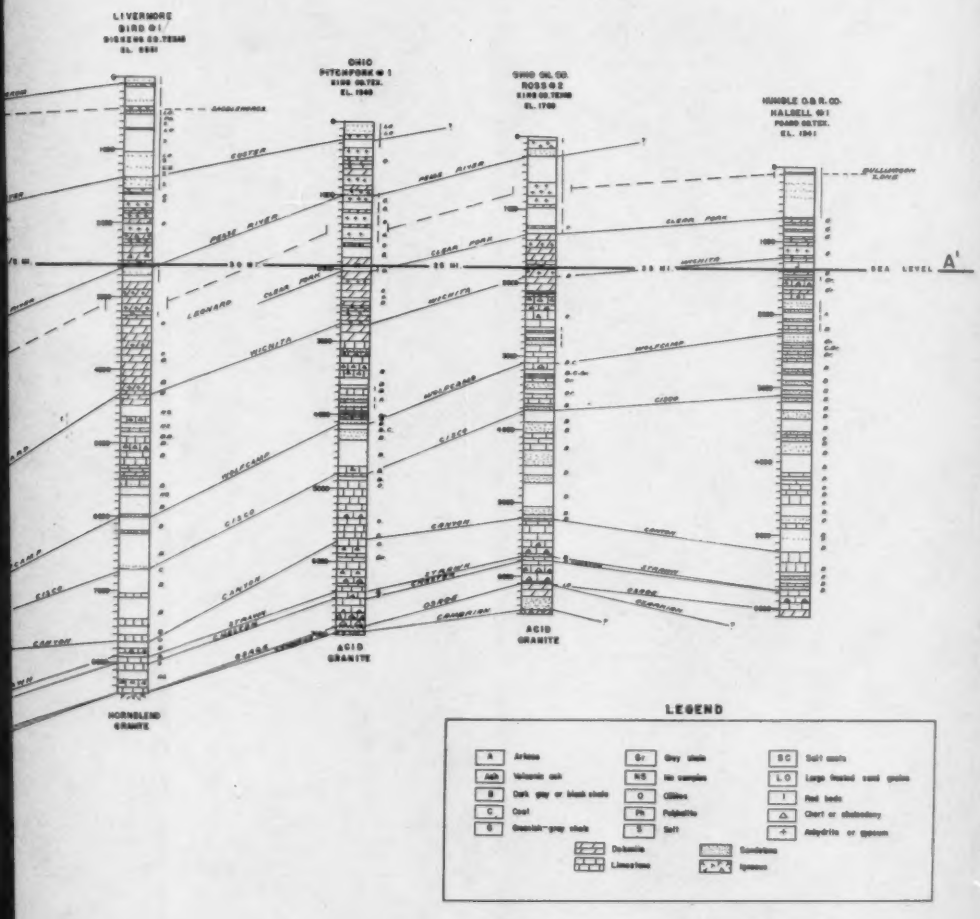
SECTION A-A'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BARNETT
MISSISSIPPIAN
DEVONIAN
ELLENBURGER
SEA LEVEL
A'
ACID GRANITE
ACID GRANITE
HORNBLEND GRANITE
LEGEND
A Arkose
Ash Volcanic ash
B Dark gray or black shale
C Coal
G Greenish-gray shale
Gr Gray shale
NS No samples
O Oolites
Ph Polyhalite
S Salt
SC Salt casts
LG Large frosted sand grains
R Red beds
Chert or chalcedony
Anhydrite or gypsum
Dolomite
Limestone
Sandstone
Igneous

SECTION A-A'

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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PALEOGEOLOGY OF THE PANHANDLE OF TEXAS

By Robert Roth

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Abstract

The sedimentary sequence and facies changes in the subsurface of the Panhandle of Texas are shown by three cross-sections. The stratigraphic column is divided into 13 geologic subdivisions. A paleogeologic map presents the areal distribution of each geologic subdivision. On each map parallel lines slanted to right indicate presence of sediments and those slanted to left indicate their absence. The effects of orogenic and epeirogenic movements are discussed.

Acknowledgment

Mr. John W. Skinner aided in the identification of fusulinids found in the cuttings from wells drilled in the area. Permission to publish has been granted by the Humble Oil and Refining Company of Houston, Texas.

Introduction

The Panhandle of Texas is situated in the southern part of the Great Plains of the United States. For many years this part of Texas has been of romantic interest because of its geographic position. As the west became settled, the Panhandle remained a comparatively undeveloped area where cattle raising was established on a large scale, but with the coming of the barbed wire fence, railroads, and small ranchers, the cattle industry became greatly restricted. The Panhandle is now a large winter wheat growing area. In 1919 natural gas was discovered on the Amarillo uplift and it has now the largest gas reservoir in the world. Petroleum is produced from a small part of the same reservoir. Because of the economic interest in the general area and because the Panhandle overlies a subsurface orogeny occurring between the Rocky Mountains to the west and the Wichita-Arbuckle Mountains to the east, it is believed that a paleogeologic study of the area would be of interest.

Nomenclature

The nomenclature used in this paper is shown by Figure 1. The more easily recognized un-

AGE	STANDARD SERIES	SERIES OR GROUPS	PANHANDLE OF TEXAS
JURASSIC	U	U	MORRISON SUNDANCE
	M	M	
	L	L	
TRIASSIC	KEUPER	DOCKUM	CHINLE / SANTA ROSA OR TRUJILLO / TECOVAS
	MUSCHELKALK	ABSENT	
	BUNTSANDSTEIN	ABSENT	
PERMIAN	GUADALUPE	CUSTER	QUARTERMASTER / WHITEHORSE / ALIBATES, TECOVAS, ROGER MILLS
	WORD	PEASE RIVER	PEASE RIVER
	LEONARD	CLEAR FORK	CLEAR FORK
		WICHITA	SCHWAGERINA crassitectoria (ZONE)
	WOLFCAMP	WOLFCAMP	WOLFCAMP
PENNSYLVANIAN — VIRGIL (CONEMAUGHIAN, MONONGAHELA)		CISCO	CISCO
PENNSYLVANIAN — MISSOURI (ALLEGHENY, CONEMAUGHIAN)		CANYON	CANYON
PENNSYLVANIAN — DES MOINES (ALLEGHENY)		STRAWN	STRAWN
PENNSYLVANIAN — DERRY (LEE, KANAWHA, POTTSVILLE)		DERRY	DERRY VERY LOCAL
PENNSYLVANIAN — MORROW (LEE, POTTSVILLE)		MORROW	
MISSISSIPPIAN	CHESTER	CHESTER	CHESTER
	MERAMEC	MERAMEC	MERAMEC
	OSAGE	OSAGE	OSAGE
	KINDERHOOK	KINDERHOOK	KINDERHOOK VERY LOCAL
SILURIAN-DEVONIAN UNDIFFERENTIATED			SILURIAN-DEVONIAN VERY LOCAL
ORDOVICIAN	CINCINNATIAN	ABSENT	
	MOHAWKIAN		
	CHAZYAN	CHAZYAN	CHAZYAN LOCAL
ORDOVICIAN?	BECKMANTOWN OR CANADIAN (ARBUCKLE, ELLENBURGER)	CANADIAN	CANADIAN LOCAL
	OZARKIAN (ARBUCKLE, ELLENBURGER)	OZARKIAN	OZARKIAN
CAMBRIAN	U	CAMBRIAN	CAMBRIAN
	M	ABSENT	
	L		

FIGURE 1.—NOMENCLATURE

conformities are at the base of the Cambrian, Mississippian, Pennsylvanian, Leonard, Custer, Dockum, and Jurassic. Evidence suggesting continuous deposition between any two series has not been recognized. Every series or group appears to be separated from those adjacent either by an unconformity or hiatus.

BASEMENT COMPLEX

The major structural features of the Panhandle of Texas (Fig. 2), control the areas of erosion and deposition. These features consist of the Anadarko geosyncline, Amarillo uplift, Plainview Basin, all asymmetrical, and the Matador archipelago or uplift. The Anadarko geosyncline is a major geologic feature extending northwestward from Oklahoma into the area. The stratigraphic dip approaches 45° along its southwest margin and many geologists believe compression faults parallel the Amarillo axis along this margin. The Amarillo uplift is an old positive line of weakness and is asymmetric in cross-section. To the southeast, the uplift comes to the surface and forms the Wichita Mountains of Oklahoma. In the subsurface, the northwestern extension of the uplift bifurcates and cannot be recognized as a unit. The asymmetric Plainview Basin, southwest of the Amarillo axis, is bounded on the south by the Matador archipelago or uplift, on the northeast by the Amarillo axis, and on the west by the Pedernal land mass of New Mexico. The Dalhart Basin is a northwestern extension of the Plainview Basin. The depositional environment of the late Pennsylvanian and early Permian in the Plainview Basin was that of a barred basin. The southern boundary of the Plainview Basin is an archipelago or uplift which greatly restricted the seaways connecting the Plainview and Midland basins; in cross-section this uplift may or may not be asymmetric. Minor structural features (Fig. 2) are generalized and their mapping may be altered both as to magnitude and trend by future drilling.

Petrographic analyses of the rocks which underlie Cambrian sediments indicate igneous types ranging from acidic to basic. They are intrusive, extrusive, and pyroclastic and represent different periods of activity. Certain areas have undergone dynamic metamorphism, others show hydrothermal alteration. Weathering of the ancient surface is slight. Igneous rocks of Wolfcamp age have been encountered in the southwestern part of the Plainview Basin. It is

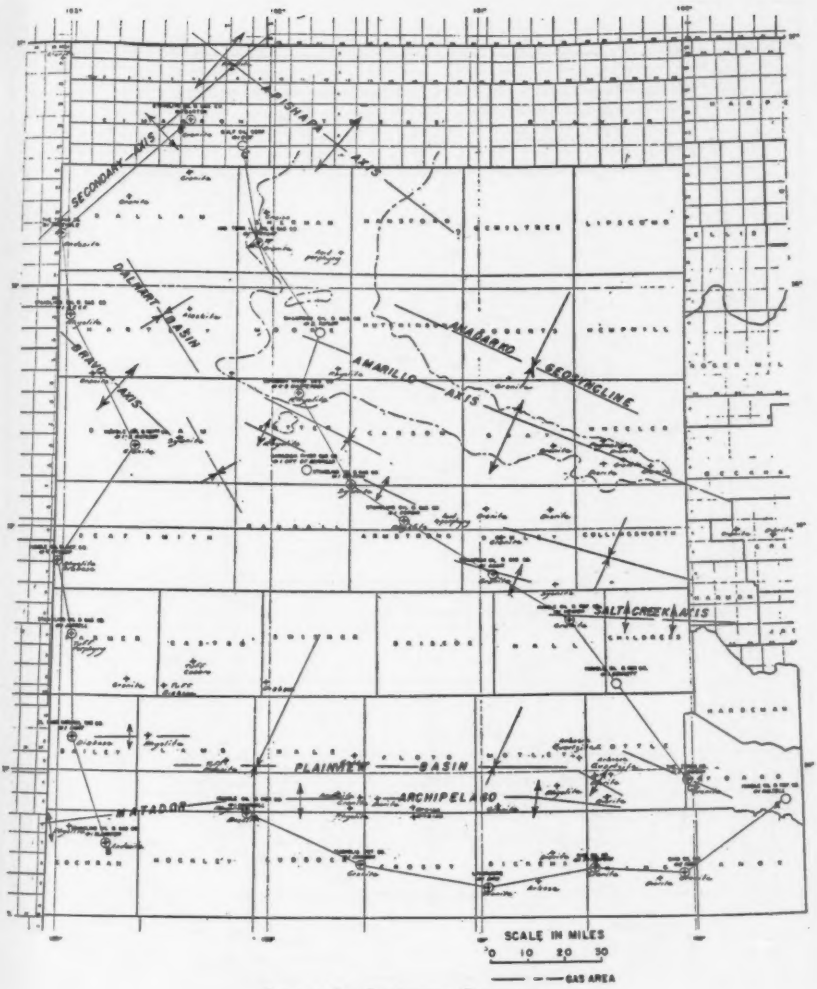

FIGURE 2.—BASEMENT COMPLEX

impossible to distinguish between Precambrian and post-Cambrian igneous rocks if Cambrian sediments are absent.

CROSS-SECTIONS

The stratigraphic sequence south of the Matador uplift, cross-section A-A′ (Pl. 1), has been influenced by the depositional environment of the Plainview and Midland basins. The abnormally thick Wolfcamp shown in the section in northern Hockley County, Texas, is due to steeply dipping sediments.

The stratigraphic sequence occurring along the western margin of the Plainview Basin, cross-section B-B′ (Pl. 2), shows the northwestward change in facies.

The stratigraphic sequence in the western part of the Anadarko geosyncline and the northeastern margin of the Plainview Basin is shown by cross-section C-C′ (Pl. 3). The line of sec-

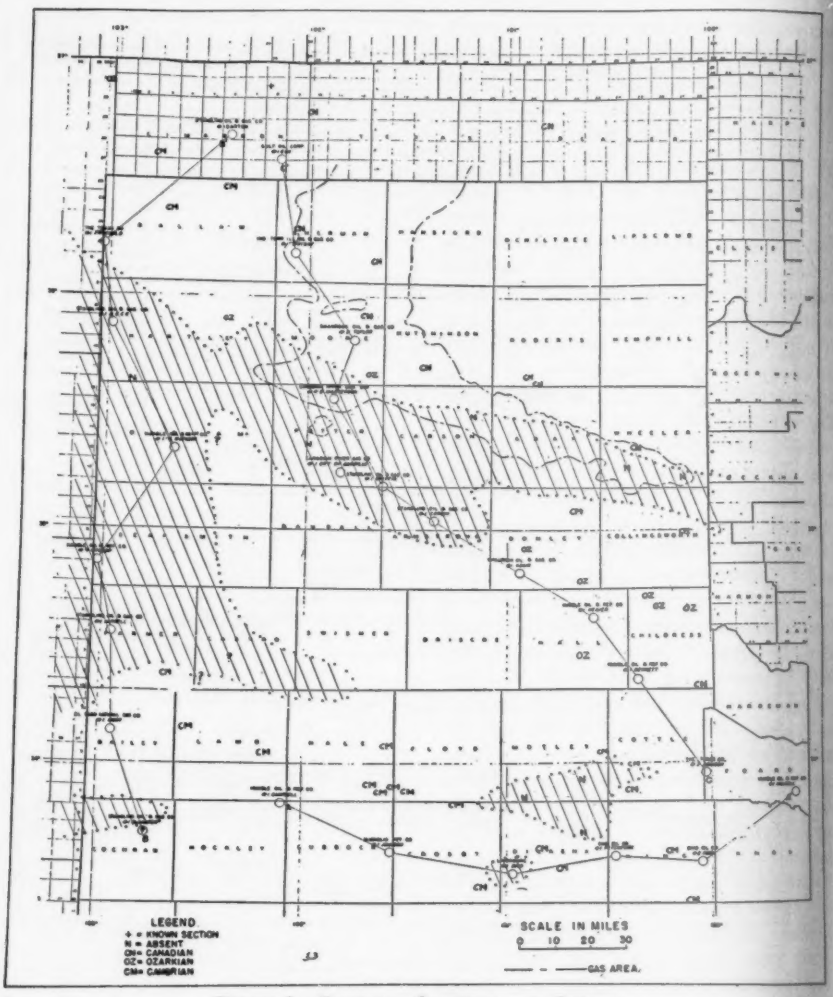

FIGURE 3.—CAMBRIAN, OZARKIAN, AND CANADIAN

tioning crosses the Amarillo axis where subsurface relief is near its maximum.

GEOLOGIC SUBDIVISIONS

Cambrian, Ozarkian, and Canadian

Cambrian sediments are coarsely crystalline dolomite, probably secondary, crystalline limestone, green shale, and large well-rounded quartz sand grains. Glauconite is common. The clastic phase of the Cambrian in King County, Texas (Fig. 3), is arkosic. Near later orogenic areas, the dolomite contains drusy vugs suggesting solution and weathering. The color of the dolomite is light gray, buff, or salmon. Because of the uniform thickness and great areal distribution of the Cambrian, it is believed that the first invasion of the Paleozoic sea was upon a peneplain. Positive areas from which sediments were derived lay outside the area. Age identification is based upon con-

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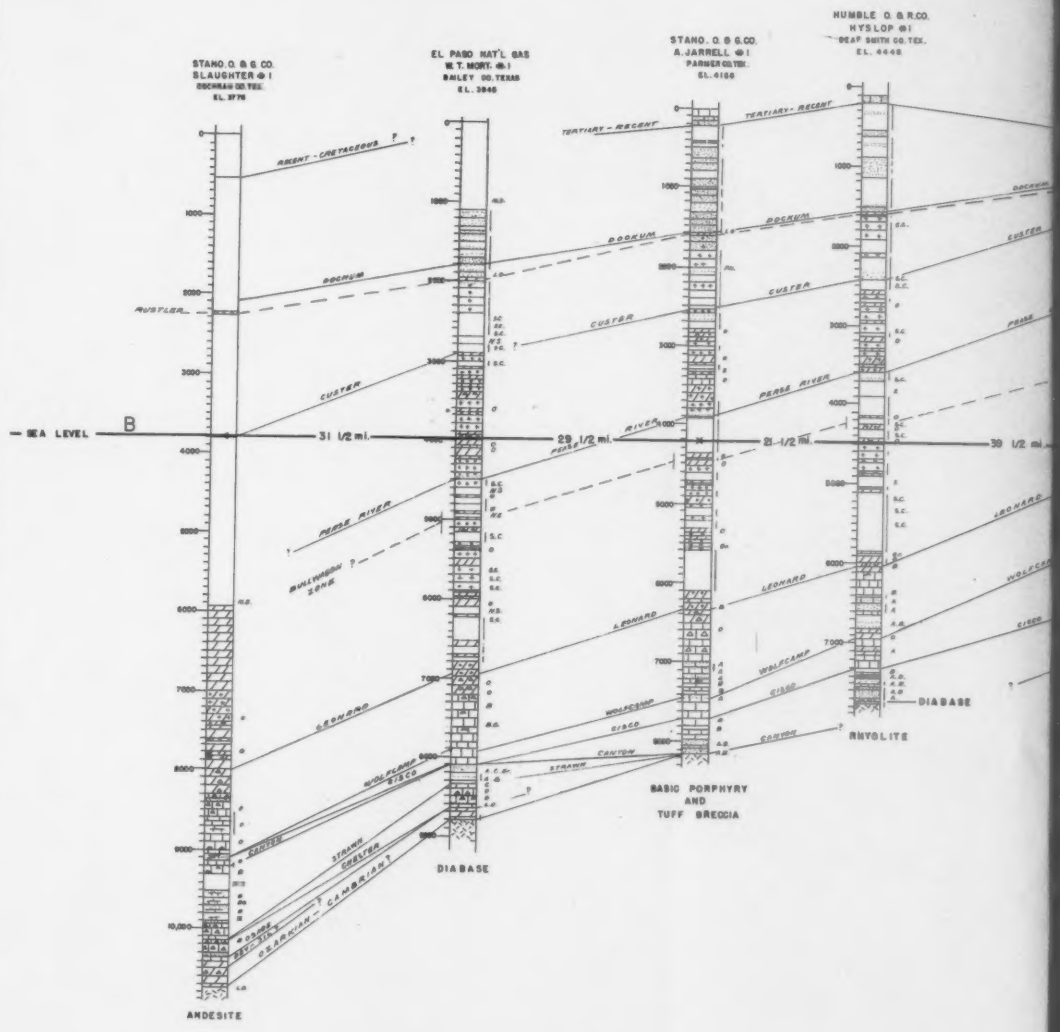
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HYSLOP #1
DEAF SMITH CO. TEX.
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30 1/2 mi.
RECENT - CRETACEOUS ?
TERTIARY - RECENT
DOCKUM
RUSTLER
CUSTER
PERSE RIVER
BALLYHOOK ZONE
LEONARD
WOLFCAMP
DISCO
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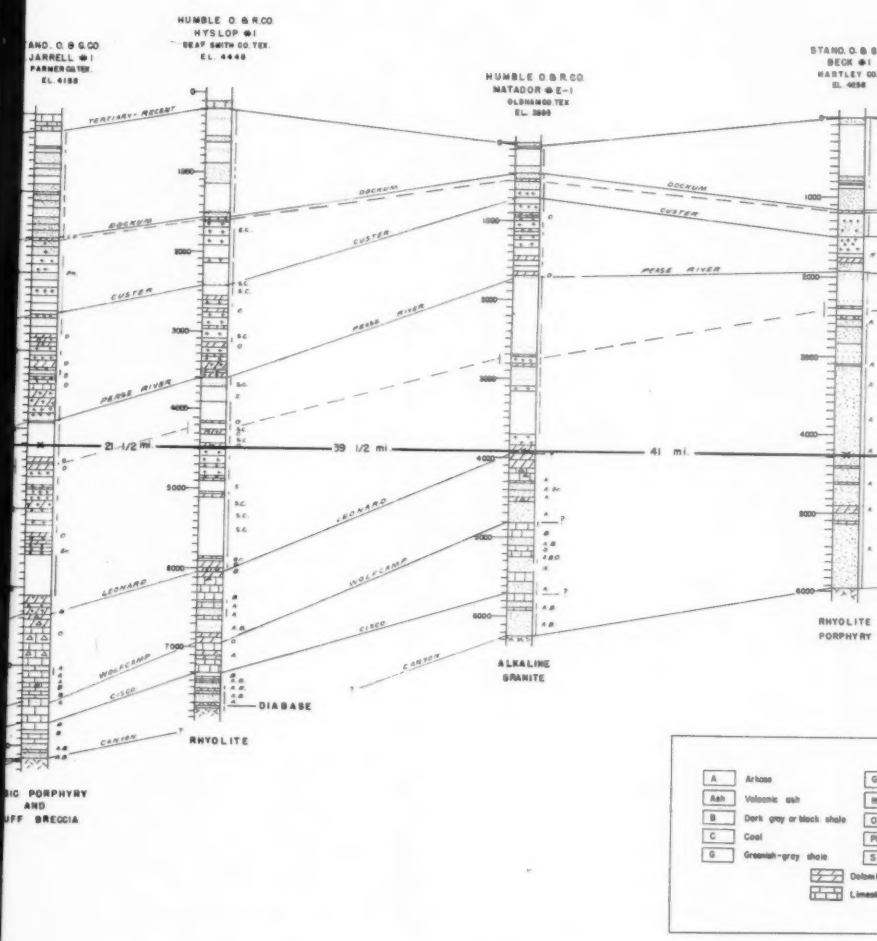
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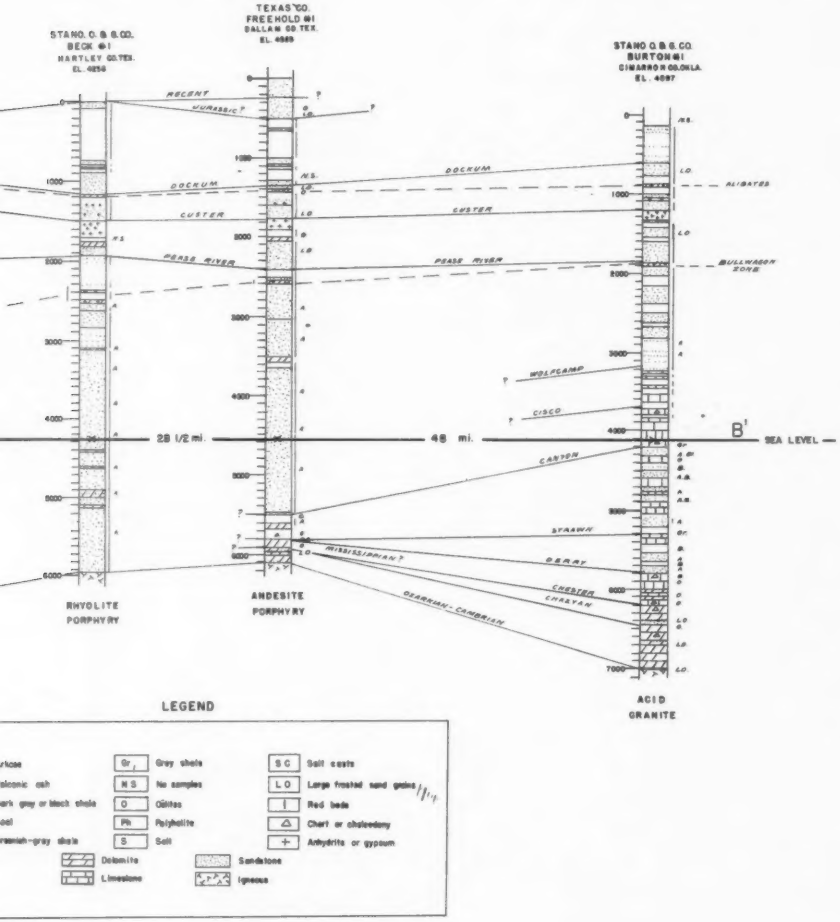
OLDHAM CO TEX
EL. 3800
TERTIARY-RECENT
DOCKUM
CUSTER
PEASE RIVER
21 1/2 mi
39 1/2 mi
41 mi.
LEONARD
WOLFCAMP
CISCO
CANYON
DIABASE
RHYOLITE
ALKALINE GRANITE
RHYOLITE PORPHYRY
A Artesia
Ash Volcanic ash
B Dark gray or black shale
C Coal
G Greenish-gray shale
Dolomi
Limest

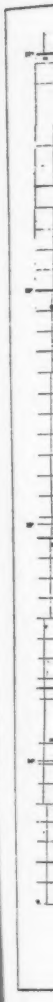
SECTION B-B'



LEGEND

Gr.	Gray shale	SC	Salt casts
NS	No samples	LO	Large frosted sand grains
O	Oolites		Red beds
Ph	Polyhalite		Chert or chalcedony
S	Salt	+	Anhydrite or gypsum

Dolomite
Limestone
Sandstone
Igneous



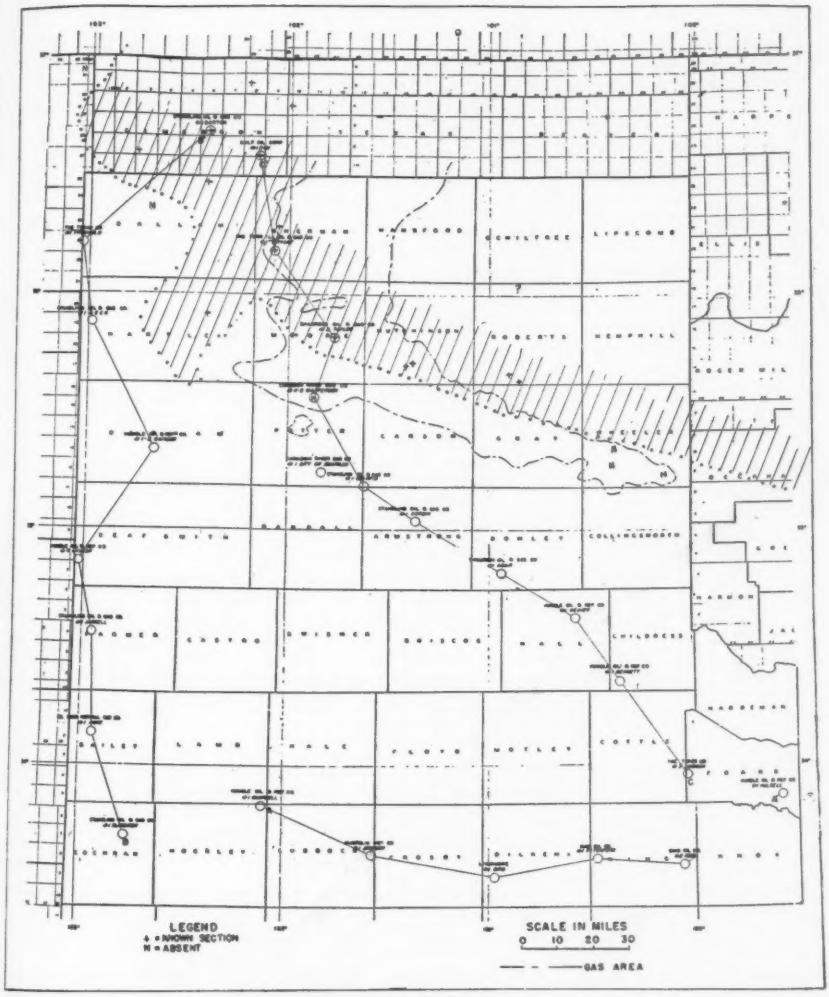


FIGURE 4.—CHAZYAN

parative lithologic characteristics and paleontology.

Ozarkian sediments are buff to salmon coarsely crystalline dolomite, probably secondary, and some variegated chert, usually oölitic. In the northwestern portion of the area (Fig. 3), a few large frosted quartz sand grains are found embedded in the dolomite. Near later orogenic areas, drusy vugs are common. The environment of the Ozarkian was similar to the Cambrian. Ozarkian orogenies have not been recognized. Age identification is based upon comparative lithologic characteristics.

Canadian sediments are dense light-gray dolomite, sometimes lithographic and calcareous, milky or translucent chert, and thin beds of bright-green shale. Some of the dolomite may be secondary. The Canadian is restricted in occurrence (Fig. 3), yet because of the absence of clastics it is believed that there were no local orogenies. Age identification is based upon comparative lithologic characteristics.

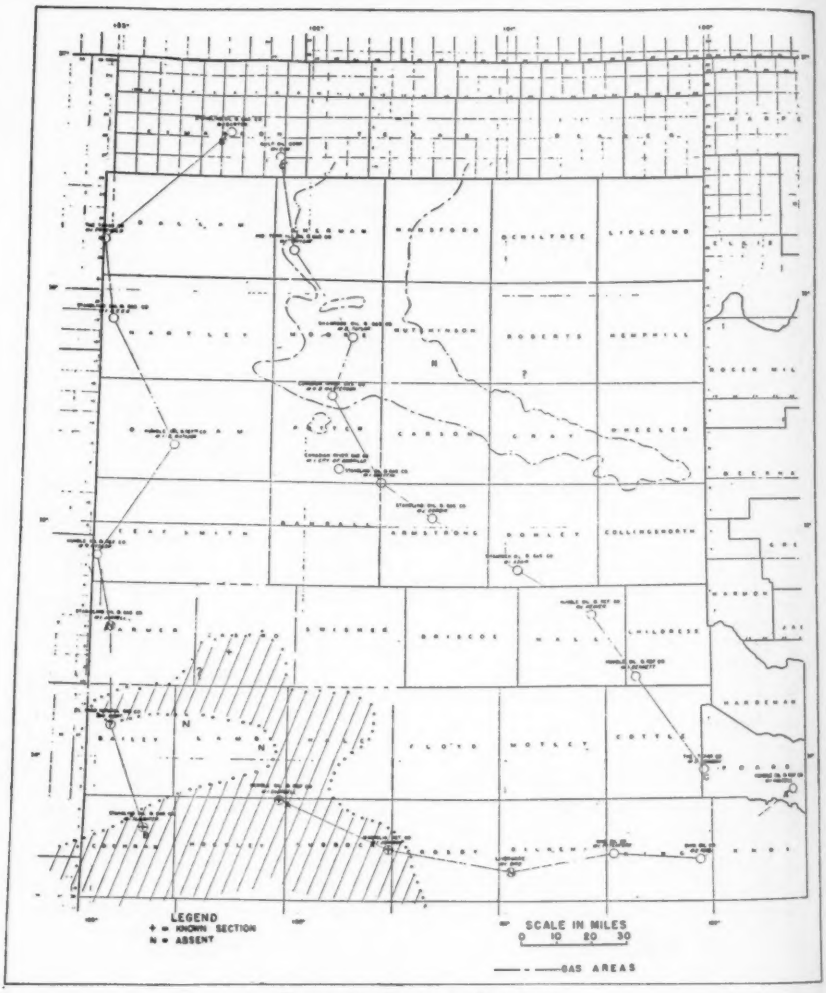


FIGURE 5.—SILURIAN—DEVONIAN

Chazyan

Sediments (Fig. 4) are very fossiliferous, coarsely crystalline, mottled light-gray limestone, locally dolomitic, coarse to fine, well-rounded frosted quartz sand grains, and bright-green shale. The sandstone contains phosphatic nodules. The Chazyan sea probably covered the entire area because lithologic sections are much the same whether encountered in the Anadarko geosyncline or in the Permian Basin. Orogenies have not been recognized. Age identification is based upon comparative lithologic characteristics and paleontology.

Silurian-Devonian

Sediments are crystalline limestone, white to light-gray dolomite, minor marl with green cast, and translucent chert. Minute grains of glauconite are common. Vugs lined with dolomite and calcite crystals in the limestone sug-

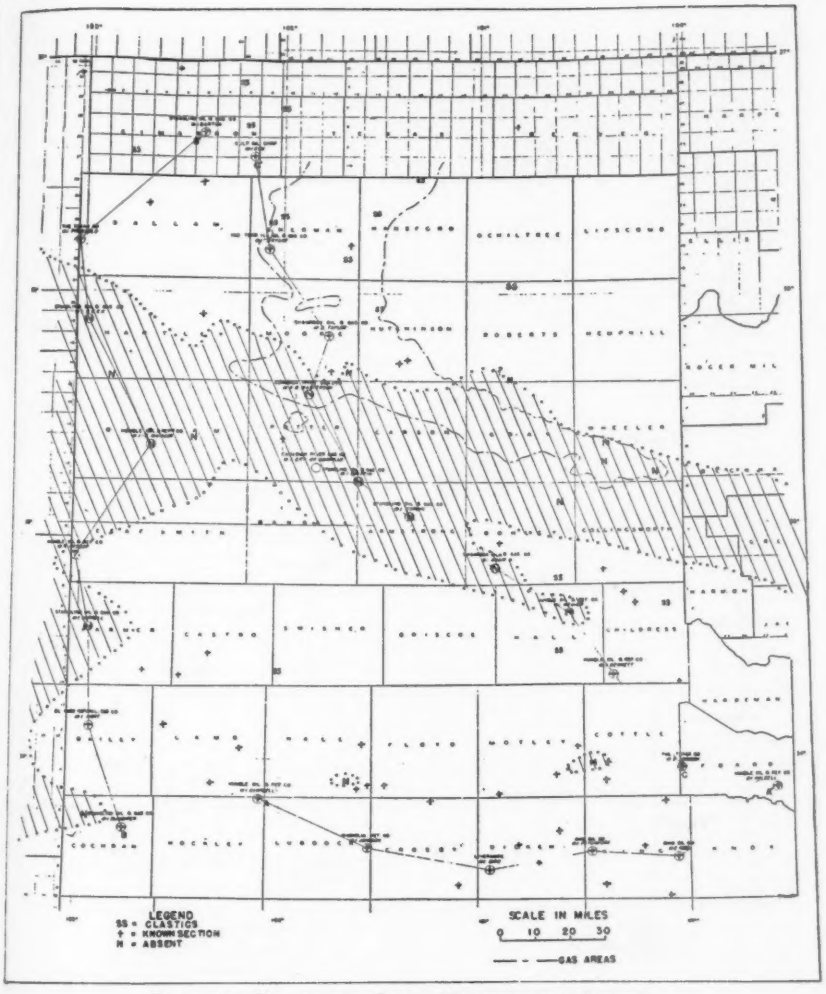


FIGURE 6.—KINDERHOOK, OSAGE, MERAMEC, AND CHESTER

gest weathering. Clastics have not been identified. The Silurian-Devonian sediments occur only in the southwest corner of the area (Fig. 5) and are a northward extension from the Permian Basin. Because of the small amount of chert present it is believed that the Devonian may be absent. The Panhandle was peneplained prior to the Mississippian invasion. Age identification is based upon comparative lithologic characteristics.

Kinderhook, Osage, Meramec, Chester

Kinderhook sediments are fossiliferous gray limestone, dark-gray chert, marl, some dolomite, and gray shale. These sediments are present in the deeper part of the Anadarko geosyncline and in the northeast part of Childress County, Texas (Fig. 6). Orogenies have not been recognized. Age identification is based upon comparative lithologic characteristics.

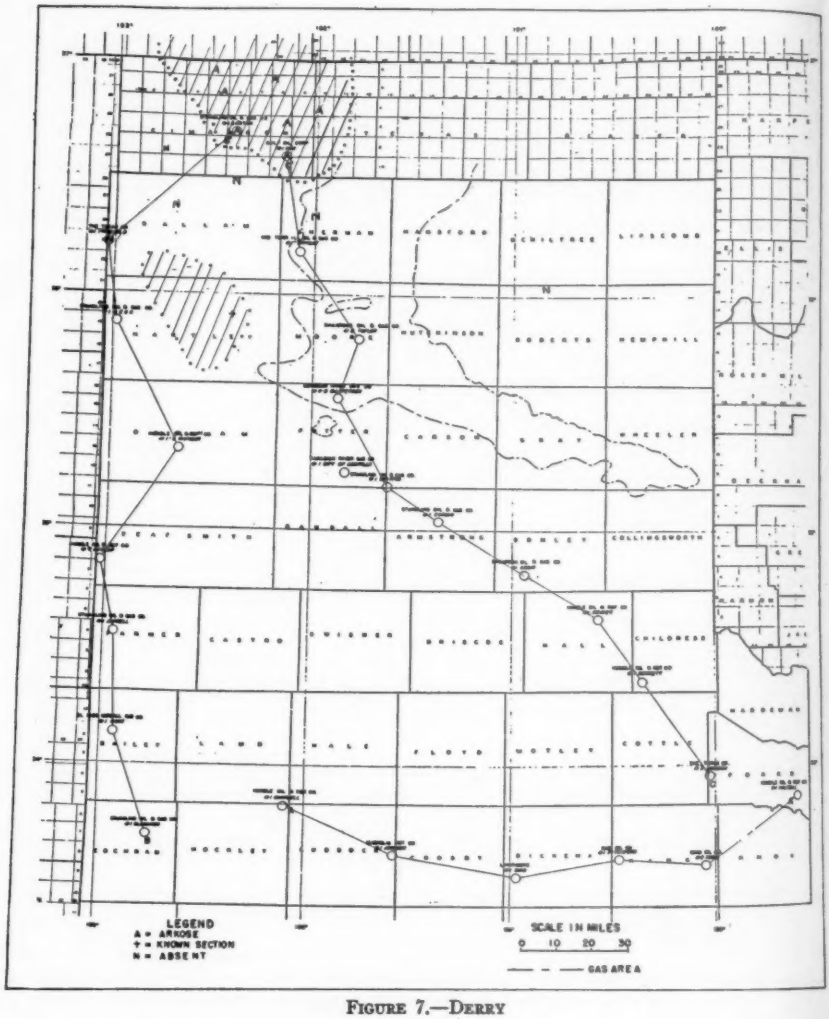


FIGURE 7.—DERRY

Osage sediments (Fig. 6) are coarsely crystalline, very fossiliferous limestone, which is very cherty in the lower part of the section. The limestone when involved in later orogenic movements is dolomitic and porous, suggesting weathering. There are no clastics. Osage sediments covered the area but have been removed from the Amarillo axis and some of the islands of the Matador archipelago. Age identification is based upon comparative lithologic characteristics and paleontology.

Meramec sediments (Fig. 6) are believed to be present in the deeper portion of the Anadarko geosyncline. However, since the western facies of the Meramec of Oklahoma is unknown and diagnostic fossils have not been found, the Meramec may be absent.

Chester sediments (Fig. 6) are white fossiliferous oölitic limestone with some chert. Along the northern and northwestern margins of the Anadarko geosyncline, and northern part of the Plainview Basin a zone of about 100 feet of fine sandstone occurs within the Chester. This clastic becomes slightly coarser toward the

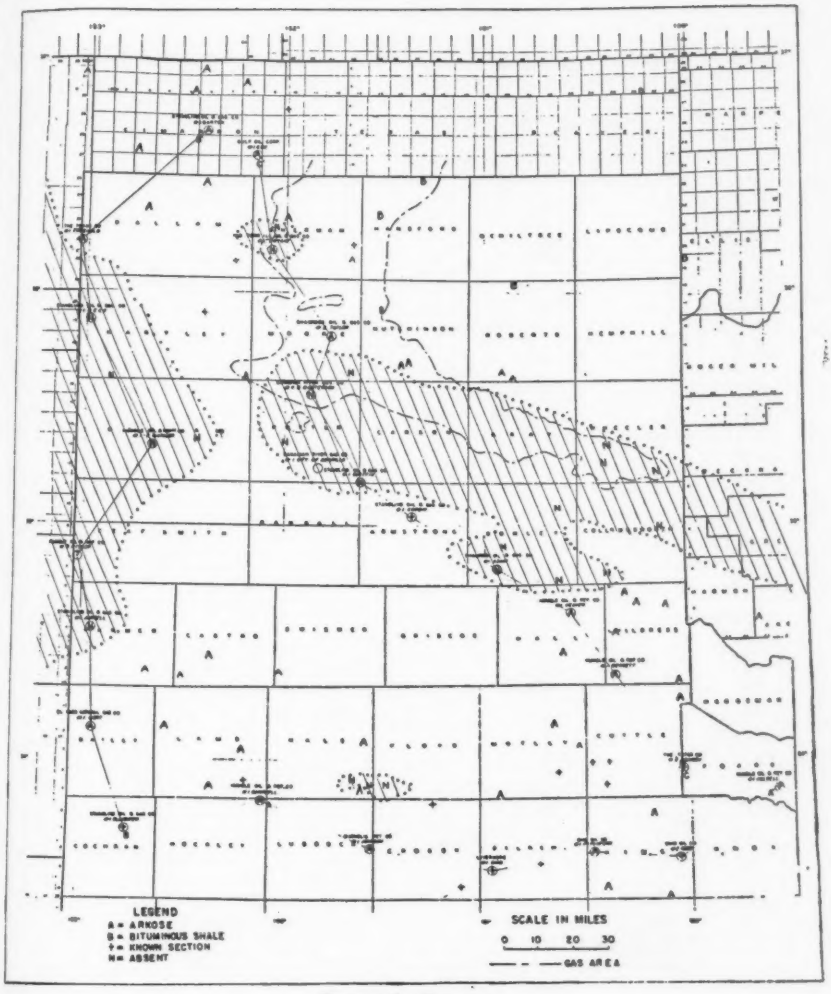


FIGURE 8.—STRAWN

northwest. A distant orogeny is suggested. Pre-Pennsylvanian erosion has removed the Mississippian from the Amarillo uplift and from most of the western part of the area. Age identification is abased upon comparative lithologic characteristics and paleontology.

Derry

Sediments (Fig. 7) are black shale, fossiliferous limestone, sometimes oölitic, quartz conglomerate, sometimes arkosic, and well cemented sandstone. Much of the sandstone, limestone, and shale contain glauconite. The presence of arkose indicates erosion of active orogenies. The stratigraphic time equivalent of the Derry may be expected in the deeper portions of the Anadarko geosyncline. Age identification is based upon comparative lithologic characteristics and paleontology. (Morrow, Atoka, and Lampasas have also been used in describing the age of the above sediments. However, the age equivalent as indicated by paleontology is nearer that of the Derry of New Mexico.)

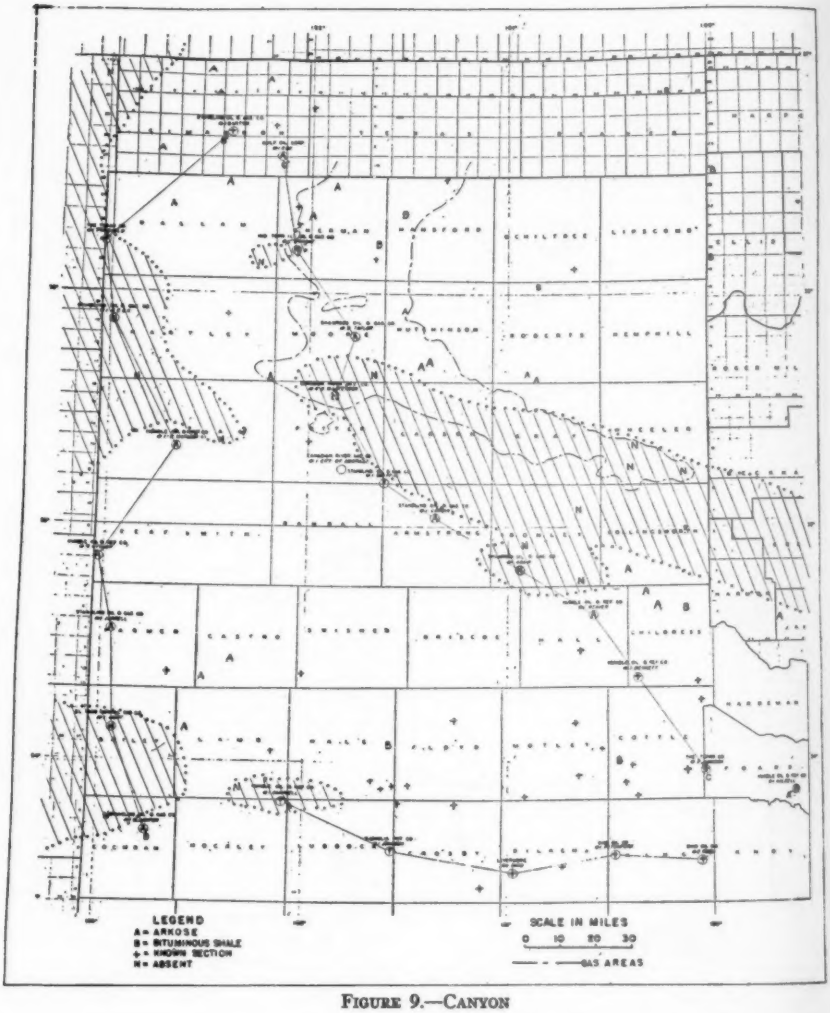


FIGURE 9.—CANYON

Strawn

Sediments (Fig. 8) are sandstone, arkose, red beds, dark-gray to black shale, coal, and fossiliferous limestone. There are some bituminous shales in the deeper parts of the Plainview Basin and Anadarko geosyncline. Red beds and arkoses increase to the west and northwest until they represent all of the Strawn in that area. The major source of the clastics was from the west and northwest; the Amarillo uplift was of secondary importance. Strawn orogenic movements established lines of weakness which persisted until the close of Wolfcamp time. A lithologic study of Strawn conglomerates adjacent to the Amarillo uplift shows that the older conglomerates are composed of fragments of dolomite and chert, and with decreasing age the conglomerates become arkoses. This indicates the removal of pre-Strawn sediments and exposure of the basement complex. One well drilled near the center of the west side of Ochiltree County, Texas, encountered anhydrite and bituminous shale. Age identification

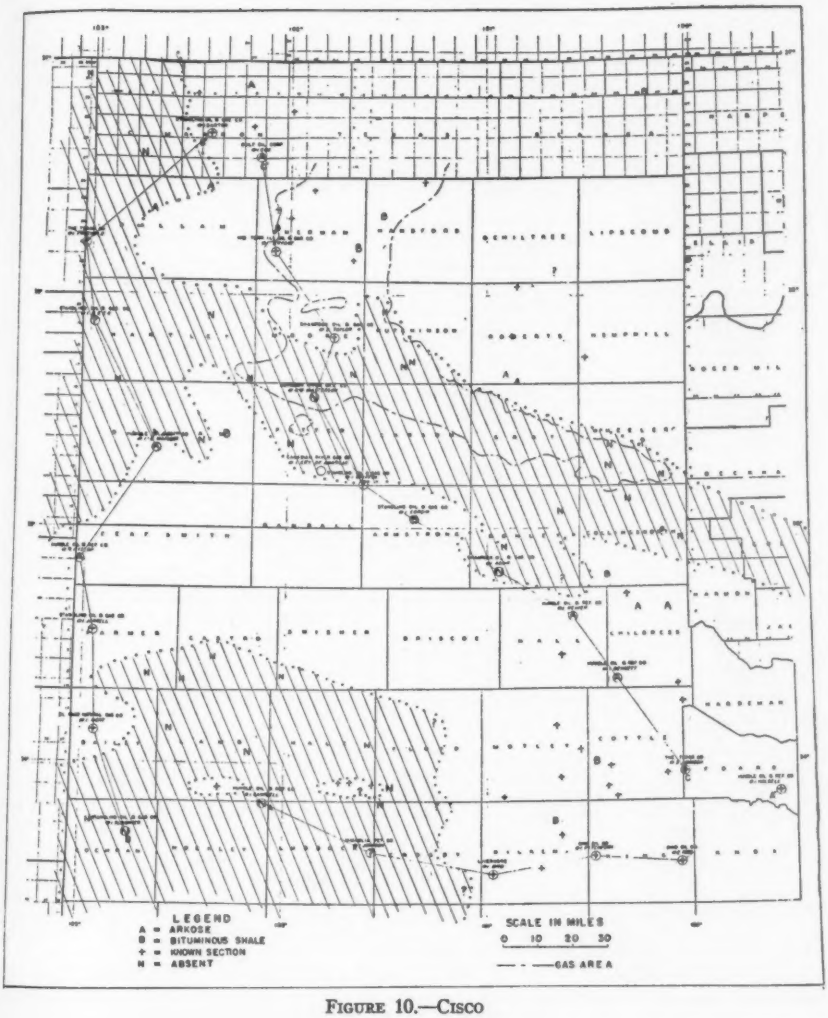


FIGURE 10.—Cisco

is based upon paleontology and comparative lithologic characteristics.

Canyon

Sediments (Fig. 9) are fossiliferous limestone, red beds, gray and dark-gray shale, coal, and coarse clastics, sometimes arkosic. Bituminous shale is present in the deeper portions of the Plainview Basin and Anadarko geosyncline. Thick deposits of limestone, becoming reef-like near the margins of positive areas, are common. Red beds and akroses increase to the west and northwest. Established orogenic weaknesses persist through Canyon time. These ended with an epeirogenic movement. Age identification is based upon paleontology.

Cisco

Sediments (Fig. 10) are red beds, gray shale, coal, coarse clastics, and fossiliferous limestone. Bituminous shale is present in the Plainview Basin and Anadarko geosyncline. Arkose is widespread and increases in thickness to the west. A greater amount of it came from the

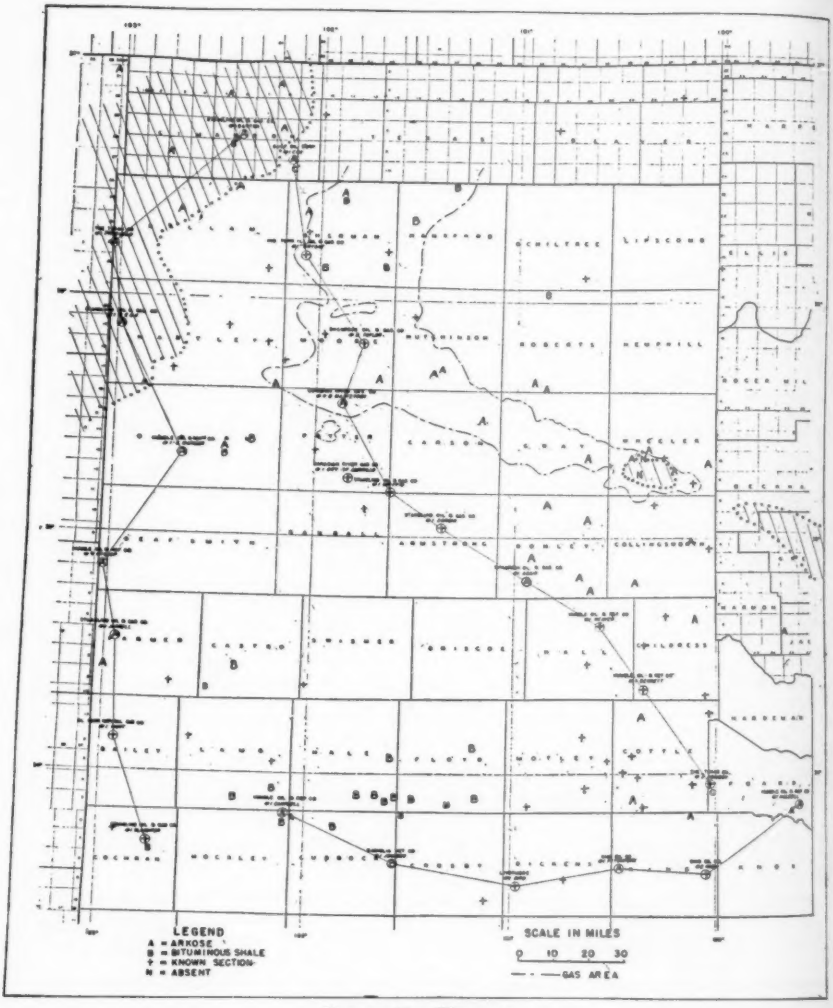


FIGURE 11.—WOLFCAMP

Amarillo uplift than previously. The Cisco sea was very shallow and reef-like masses of limestone were built upon local highs, such as the Matador archipelago and bordering areas of the Amarillo uplift. In northeast Dickens County, Texas, anhydrite was deposited with bituminous shale. Age identification is based upon paleontology.

Wolfcamp

Sediments (Fig. 11) are fossiliferous limestone, porous secondary dolomite, chert, chalcedony, variegated shale, and coarse clastics. Bituminous shale was deposited in the deeper portions of the Plainview Basin and Anadarko geosyncline. To the northwest, the Wolfcamp grades into red beds and arkose. A local source of arkose was the Amarillo uplift. Epeirogenic uplift ended Wolfcamp deposition. Age identification is based upon paleontology and comparative lithologic characteristics.

Wichita-Clear Fork

Wichita sediments (Fig. 12) are red beds, evaporites, primary and secondary dolomite,

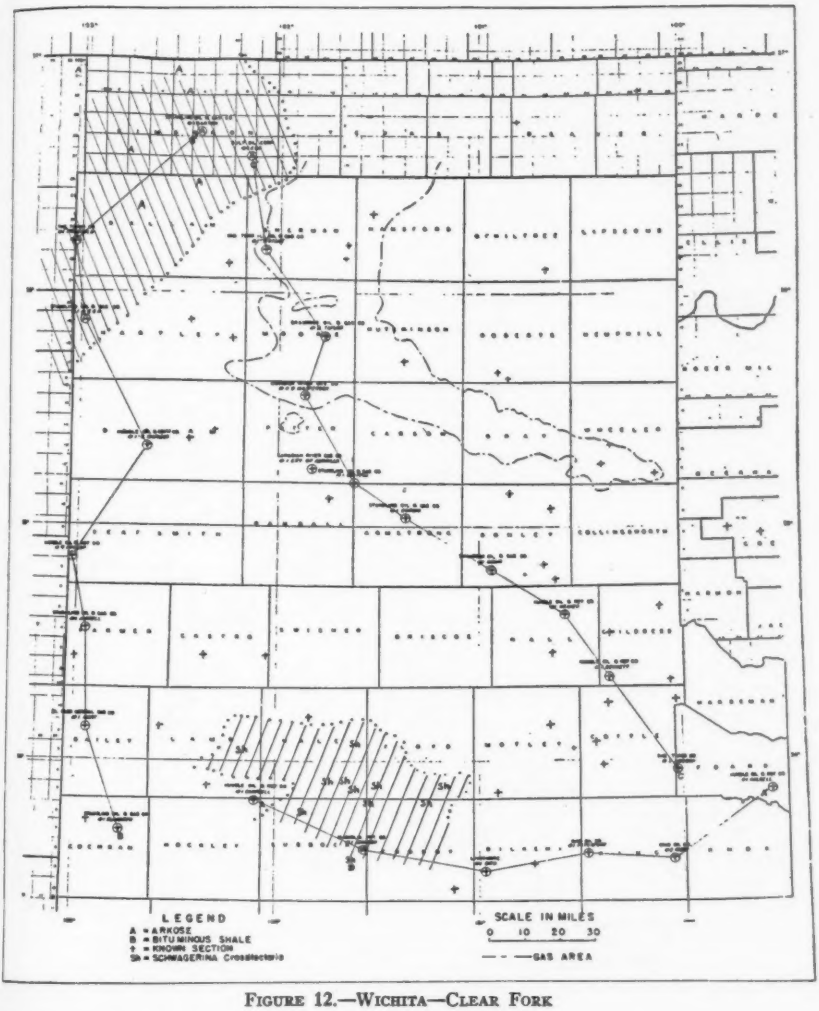


FIGURE 12.—WICHITA—CLEAR FORK

fossiliferous limestone, and bituminous shale. To the northwest where Wichita sediments are relatively thin, they are mainly red beds and arkoses with minor evaporites. In the southern portion of the area, they are dolomite, limestone and bituminous shale. The Wichita is the oldest formation now covering the Panhandle of Texas. Age identification is based upon comparative lithologic characteristics and paleontology.

Clear Fork sediments (Fig. 12) are red beds, evaporites, primary and secondary dolomite, and arkose. The dolomite becomes very oölitic adjacent to the Matador archipelago. Arkose is present only in the northwestern part of the area. Clear Fork sediments become increasingly red in a northerly direction. This facies change is due to negative warping and to a southward retreating sea. Age identification is based upon comparative lithologic characteristics in the subsurface and invertebrate paleontology where the formations crop out.

Pease River-Custer

Pease River sediments (Fig. 13) are evaporites; red and greenish-gray shale; primary and

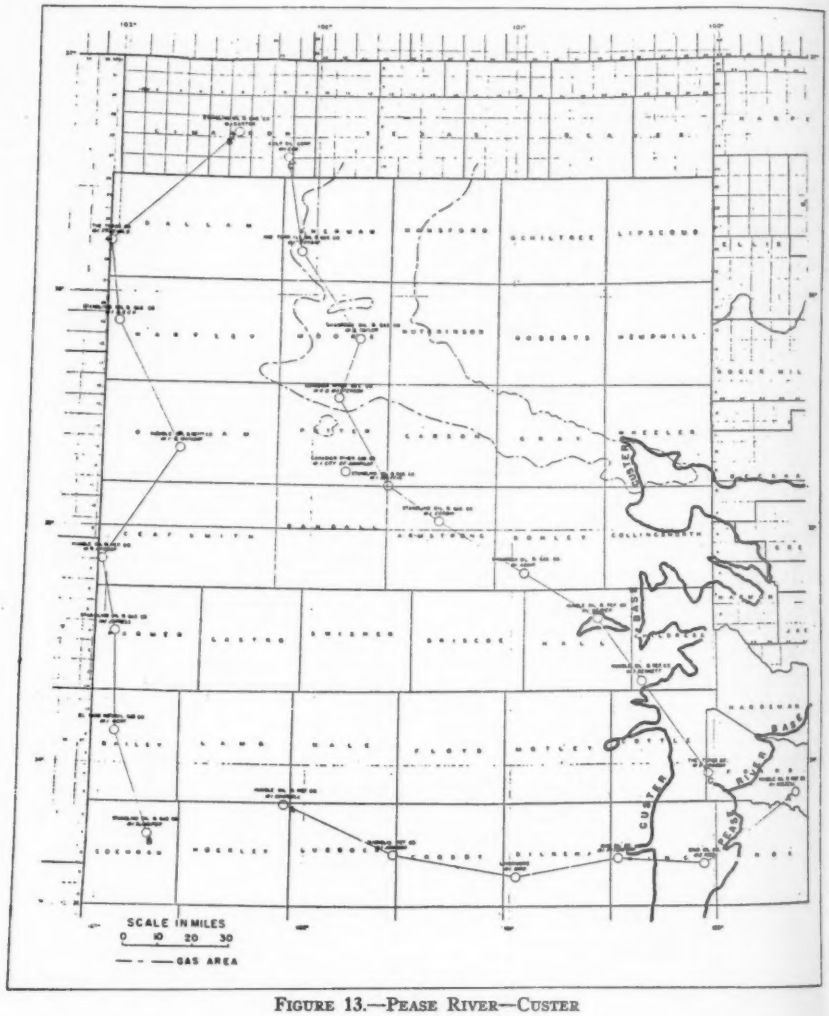


FIGURE 13.—PEASE RIVER—CUSTER

secondary dolomites, some oölitic; limestone, usually oölitic; coarse clastics; and volcanic ash. A basal sandstone bed, less than 100 feet thick and with indistinct upper contact, is very widespread. Its sand grains are fine and angular with very few large frosted grains. Northwest of Moore County, Texas, this basal sandstone thickens to approximately 400 feet. Its sand grains are coarse, rounded, polished, and when not leached have a bright-orange color. Epeirogeny, with no marked orogeny, ended Pease River deposition. Age identification in the subsurface is by comparative lithologic characteristics and where the formations crop out by invertebrate paleontology.

Custer sediments (Fig. 13) are evaporites including magnesium salts, terra cotta shale, primary and secondary dolomite, volcanic ash, chert, and coarse clastics. These coarse clastics include fine orange polished sand grains with zones of large white or sometimes orange-red frosted sand grains. Petrographic analyses of

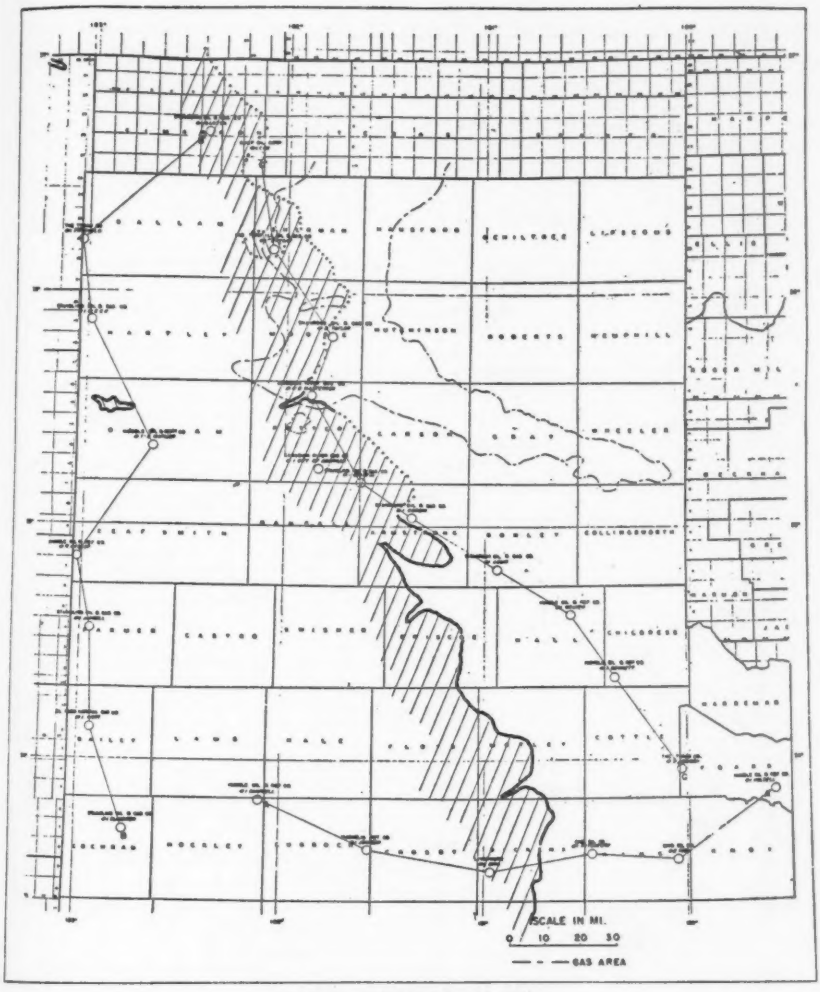


FIGURE 14.—DOCKUM

some of the sandstones show that they were derived in part from pegmatitic granites. The Pease River-Custer contact is marked by a hiatus. Regional tilting southward but no orogenies occurred during Custer time. Age identification in the subsurface is based upon comparative lithologic characteristics and where the formations crop out by invertebrate paleontology.

Dockum

Sediments (Fig. 14) are variegated shale and coarse clastics with some volcanic ash as cementing material. The conglomerates present in the northern part of the area were derived from older sedimentary beds and are usually called "clay-ball", while those present in the southern part of the area were derived from a Precambrian metamorphic complex. Marine

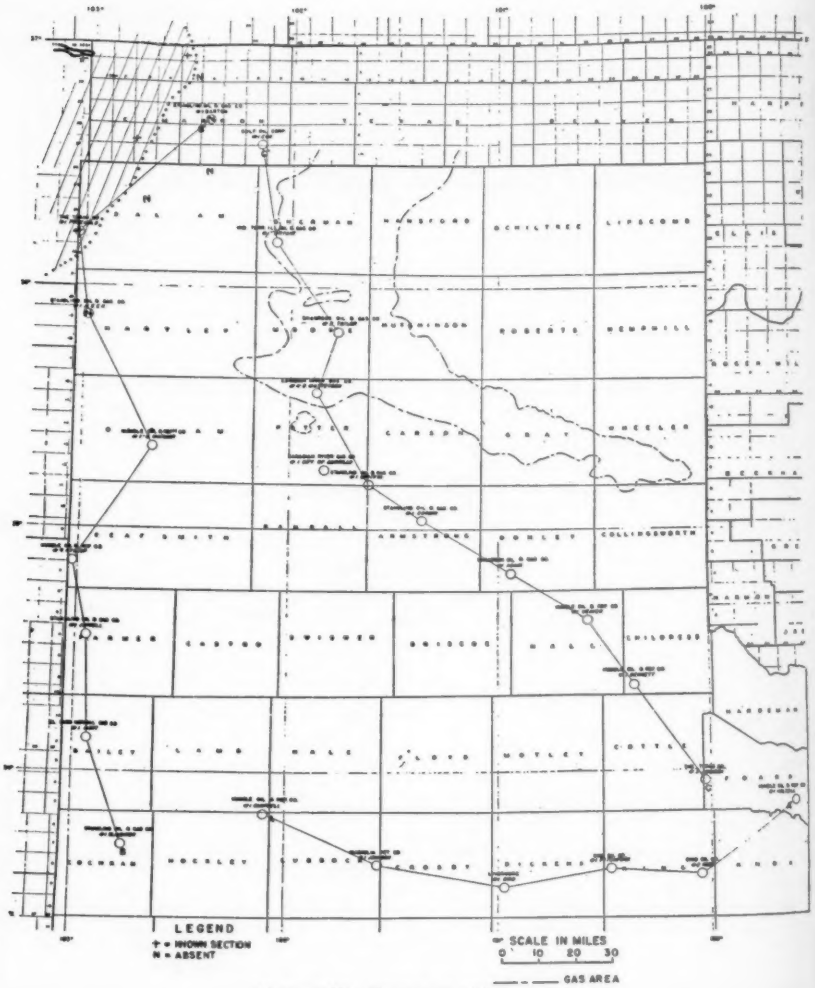


FIGURE 15.—SUNDANCE—MORRISON

Upper Triassic fossils have not been found. A marked lithologic disconformity separates the Triassic from the Permian and indicates widespread epeirogenic movements of pre-Upper Triassic age. Age identification in the subsurface is based upon comparative lithologic characteristics, and, where formations crop out, vertebrate paleontology and paleobotany.

Sundance-Morrison

Sediments (Fig. 15) are light-gray, green, and maroon shale with coarse clastics. A thick bed of sandstone occurs at the base of the series and much of its sand grains are well-rounded and frosted. In Cimarron Canyon, Union County, New Mexico, the Jurassic rests with angular unconformity upon older beds indicating pre-Jurassic orogenic movements. The Jurassic has been tilted by the Rocky Mountain uplift. Age identification is based upon comparative lithologic characteristics in the subsurface, and vertebrate paleontology where the formations crop out.

Post-Jurassic

Sediments of Comanchean age are widely scattered over the area. The "cap rock", of Tertiary age, is composed of caliche and clastics and covers a large portion of the Panhandle of Texas. "Blow sand", of Recent age occurs along the rivers. Except for a widespread invasion by the Comanchean sea, the Panhandle of Texas has been a quiescent land mass undergoing peneplanation since Triassic time.

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Manuscript Received by the Secretary of the Society, November 22, 1948

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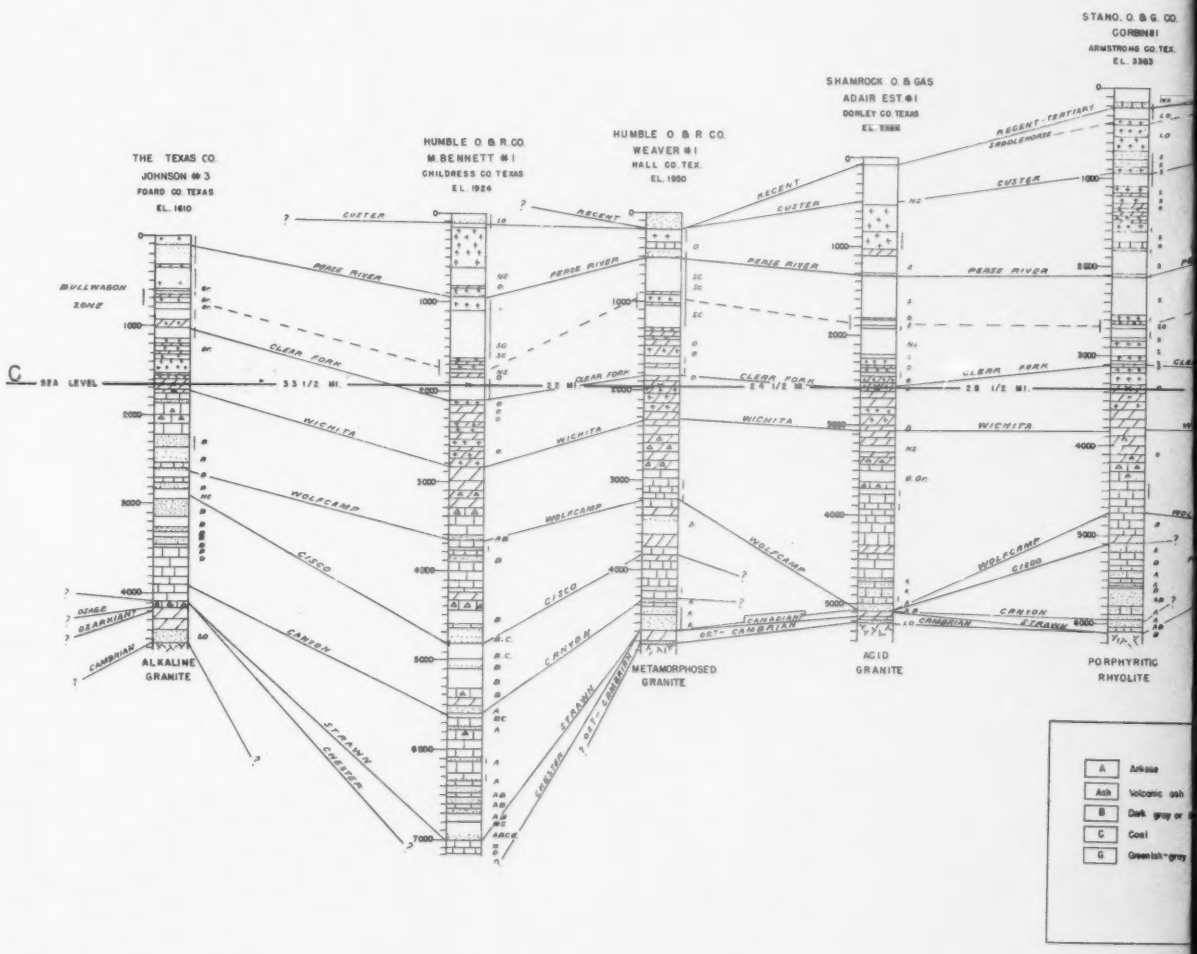
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WOLFCAMP
CISCO
CANYON
STRAWN
CHESTER
ORT.-CAMBRIAN
CANADIAN
CAMBRIAN
OSAGE
DEARMANIAN
ALKALINE GRANITE
METAMORPHOSED GRANITE
ACID GRANITE
PORPHYRITIC RHYOLITE
A Arkose
Ash Volcanic ash
B Dark gray or
C Coal
G Greenish-gray

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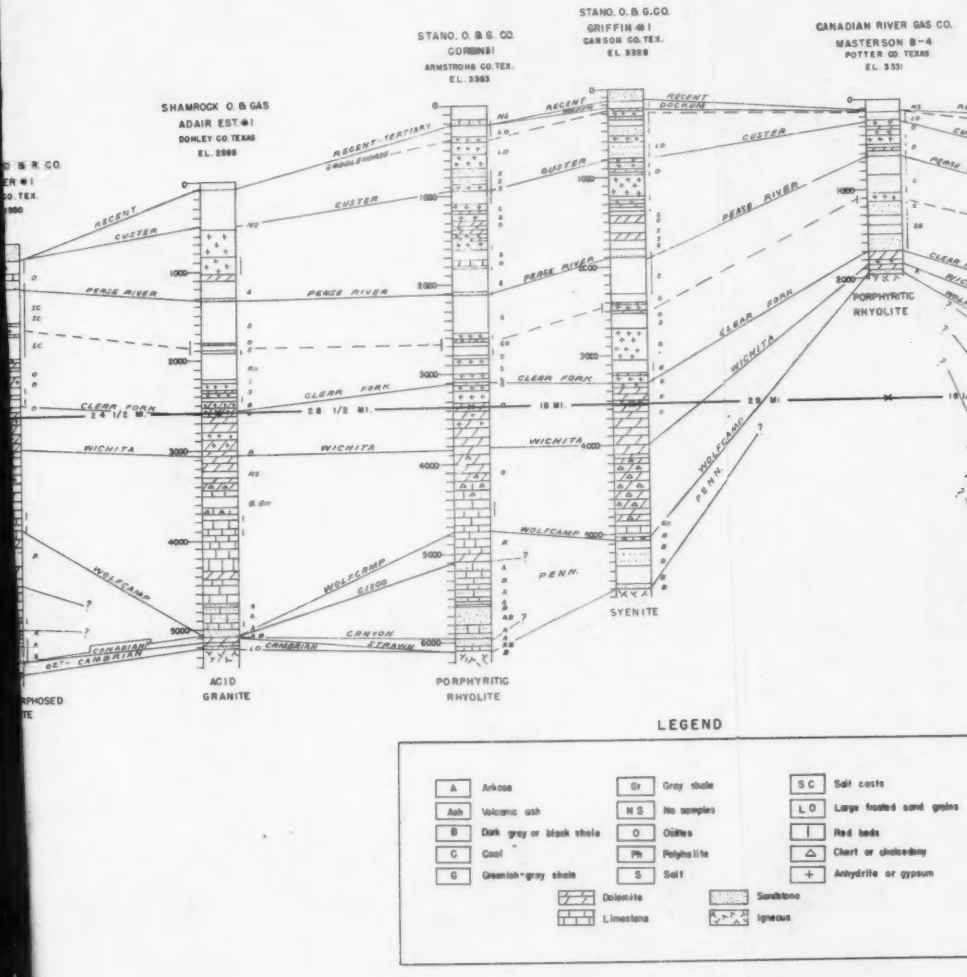
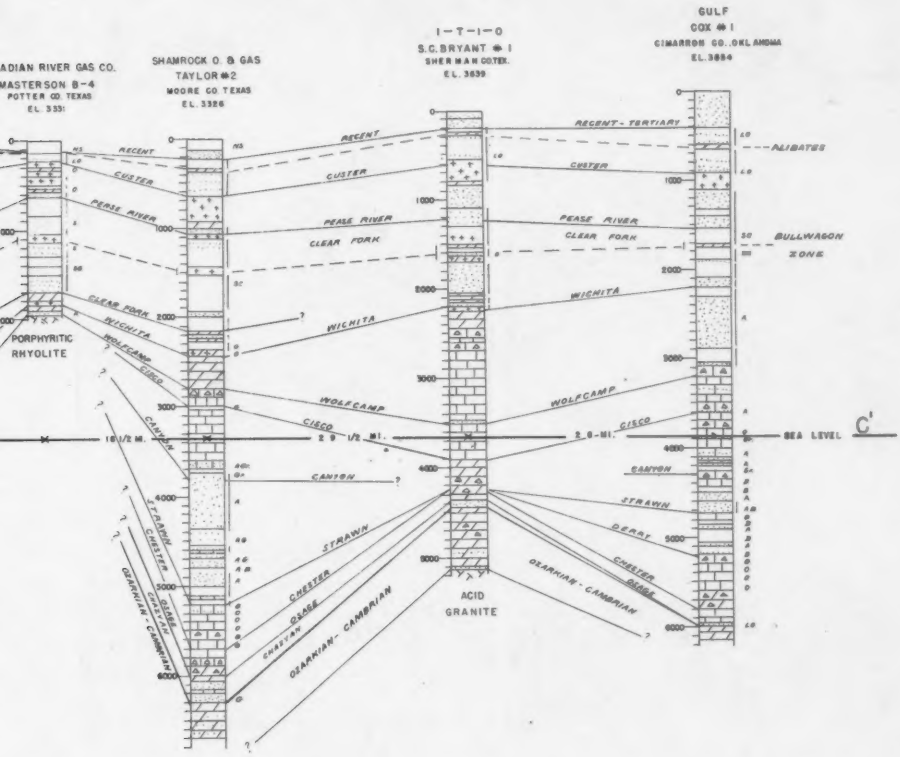


LEGEND

A Arkose	Gr Gray shale	S C Salt casts
Ash Volcanic ash	N S No samples	L O Large frosted sand grains
B Dark gray or black shale	O Oolites	Red beds
C Coal	Ph Polyhalite	Chert or chalcedony
G Greenish-gray shale	S Salt	+ Anhydrite or gypsum
Dolomite	Sandstone
Limestone	Igneous

SECTION C-C'



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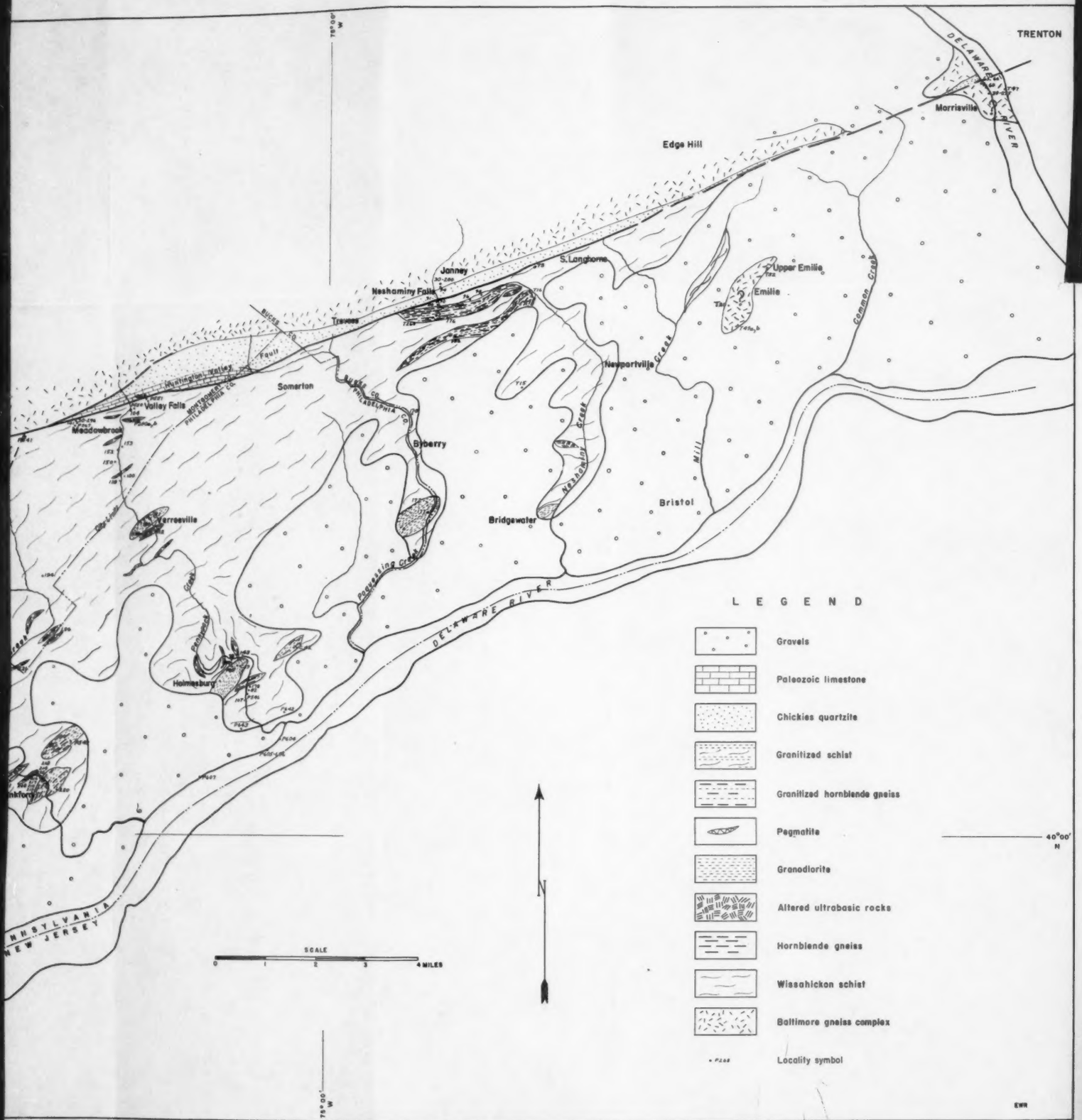
WISSAHICKON SCHIST AT PHILADELPHIA, SHOWING AREAL GEOLOGY, METAMOR



G AREAL GEOLOGY, METAMORPHIC ZONES, AND NUMBERED LOCALITIES

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WISSAHICKON SCHIST AT PHILADELPHIA, PENNSYLVANIA

By Judith Weiss

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Abstract

The Wissahickon schist of the region in and near Philadelphia originally was a sediment made up of sandy, shaly, and arkosic layers of varying thickness and composition. These sediments have been completely recrystallized and highly deformed during regional metamorphism; however, the original com-

positional banding has been retained. The presence of the metamorphic index minerals, garnet, staurolite, kyanite, and sillimanite, in definite zonal arrangement, indicates that the regional metamorphism increased in intensity from northwest to southeast. The rocks of the area belong to the amphibolite and epidote-amphibolite facies.

Granitization along the southwestern border of the area has altered the schist. Intrusive basic igneous rocks in the southern and eastern part of the area also show some granitization. Apart from the granitized rocks the chemical composition of the schist has not been changed appreciably, and the metamorphism may be considered as isochemical.

The structure of the rocks is believed to be that of an abnormal synclinorium produced by compression. No explanation is offered for the coincidence of the northern limb of the synclinorium with the garnet and staurolite zones, the center with the kyanite zone, and the southern limb with the sillimanite zone. The northern limb of the synclinorium and the zones of regional metamorphism are truncated by faults.

Acknowledgments

Gratitude is extended to Dr. Edward H. Watson of the Department of Geology of Bryn Mawr College, for his suggestion of the problem and for his interested and helpful direction of research. The author is grateful to Dr. Lincoln Dryden of the Department of Geology of Bryn Mawr College, for his suggestions and proofreading of the manuscript. The author wishes to thank both Dr. Ernst Cloos of the Department of Geology of Johns Hopkins University and Dr. Earl Ingerson of the United States Geological Survey for their critical reading of the paper.

Introduction

Problem and Area

This paper deals with the structure, petrology, and metamorphism of the Wissahickon schist in and near the city of Philadelphia, Pennsylvania. The Wissahickon schist was named for Wissahickon Creek in Philadelphia. Subsequently, the name was applied to rocks of similar appearance farther to the southwest in Pennsylvania, Maryland, and Virginia. The purpose of the study is to record and correlate the structure, petrology, and metamorphic zones of this type area.

The area under consideration, 35 miles long and 8 miles wide, extends from Morrisville, Penna., on the Delaware River, west to Darby Creek in Delaware County. Paleozoic rocks and Baltimore gneiss form the northern boundary and the Delaware River is the southern boundary. The recently studied granodiorites (Postel, 1940) west of the Schuylkill River and south of City Line are not included. The area covers parts of Burlington, Philadelphia, Germantown, Norristown, and Chester quadrangles (Pl. I).

In this paper the formation is called Wissahickon schist in order to avoid the cumbersome phrase "schist and gneiss", and because the most prevalent structure is schistose. The name applies to those recrystallized dominantly pelitic rocks that are south of the Cream Valley-Huntingdon Valley fault (Bascom, 1909a, p. 16) and Rosemont fault (Armstrong, 1941, p. 684), and that extend southwest into Delaware and Chester counties.

Methods of Work

About 130 days were spent in the field and twice that number in the laboratory. Eighty-eight new thin-sections were cut and 100 more from previous collections were studied. The modes of 103 thin-sections were determined with an integrating stage.

In order to determine the occurrence of index minerals in the various zones of metamorphism, 84 heavy mineral separations were made with Thoulet's solution and 25 with methylene iodide. Fourteen additional separations were obtained from L. Dryden of the Department of Geology at Bryn Mawr and 10 from Jane Bell formerly of Bryn Mawr Geology Department.

Three new chemical analyses were made by members of the chemistry department of Bryn Mawr College; nine other analyses were obtained from previously published papers and from the unpublished notes of Bascom, for many years head of the Department of Geology of Bryn Mawr College.

General Geology

Coastal Plain deposits of unconsolidated clays, sands, and gravels of Cretaceous, Tertiary, and Quaternary age overlap the Wissahickon schist to the southeast of the area

studied. The overlap extends west of the Delaware River (Pl. 1). The schist is exposed along the major streams and at altitudes above 100 feet, but some small, isolated patches of gravels cover the schist up to altitudes as high as 400 feet.

On the north and northwest, the Cream Valley-Huntingdon Valley and Rosemont faults separate the schist from Precambrian Baltimore gneiss and Paleozoic sediments. These sediments are Cambro-Ordovician limestone that underlies Huntingdon Valley and Cambrian Chickies quartzite that crops out between Somerton and Edge Hill. The Baltimore gneiss, a composite rock largely of igneous origin, is older than the Wissahickon schist (Bascom, 1909a, p. 3). However, since in this region a high-angle thrust fault separates the schist from the northern bordering rocks, the stratigraphic relation between the schist and the Paleozoic sediments is not clear.

West and southwest of Philadelphia, the schist has been extensively granitized and also intruded by granodiorites (Postel, 1940; Bascom, 1909a, p. 5–6). Within the area of this study, small granodiorite bodies occur, but no extensive granitization was noted. There are igneous intrusions in the schist of numerous sheets and lenses of hornblende gneiss and a small number of ultrabasic rocks

Previous Work

The rocks of the Philadelphia area have been studied for almost 100 years. The lithological variation of the schist was noted long before 1900, and many field observations of early geologists can be correlated with the results of modern petrographic work.

Rogers (1858) spoke of the mica slate along Wissahickon Creek Valley. Lesley (1892, p. 118–128) called the rocks the Newer Gneiss and divided the section along the Schuylkill River into the Philadelphia Schist and Gneiss, the Manayunk Schist, and the Chestnut Hill Garnet Schist Group.

Rand (1901) accepted Lesley's Chestnut Hill Garnet Schist Group and Manayunk Schist, but divided the Philadelphia Group into the Porphyritic Gneiss, the Fairmount Gneiss, and the Frankford Gneiss. He also added another type, the Spangled Schists, that represent the Cream Valley-Huntingdon Valley fault zone. The Chestnut Hill Schists correspond to the rocks of the garnet and staurolite zones of this paper. The Manayunk Gneisses and Schist represent the schist of the sillimanite zone exposed along the Schuylkill River. The Porphyritic Gneiss, called by Rand an augen-gneiss, is the migmatite or granitized schist in the vicinity of West Laurel Hill (Postel, 1940). The Fairmount Gneiss, described as not extensive and composed of fine-grained muscovite-microcline gneiss, is that part of the granitized schist which occurs between City Line and 3 miles south of City Line on the east side of the Schuylkill River. The Frankford Gneiss refers to bodies of granitized hornblende gneiss and possibly the now hidden granodiorite in Frankford.

Bascom (1902) assigned the name Wissahickon to the formation because of the excellent exposures along Wissahickon Creek Valley. Mathews (1904) divided the formation into Wissahickon phyllite, Wissahickon mica schist, and Wissahickon mica gneiss. He believed the rocks were Ordovician. Bascom (1905), in describing the Piedmont of Pennsylvania, also called the Wissahickon mica schist and mica gneiss Ordovician but suggested a possibility of a Precambrian age for the Wissahickon mica gneiss. In the Philadelphia Folio, Bascom (1909a, p. 4) assigned the Wissahickon mica-gneiss, which includes the rocks studied in this paper, to the Precambrian.

Many geologists continued to study the Wissahickon schist (Bliss and Jonas, 1916; Knopf and Jonas, 1923; Hawkins, 1924; Miller, 1935; Mackin, 1935). There was much emphasis on the age problem and stratigraphic relationships of the formations in the Pennsylvania Piedmont. However, detailed petrographic and structural study and work such as that of Cloos and Hietanen (1941) must be made of the entire formation, together with work in the Paleozoic section, before the problem of the age of the Wissahickon schist can be solved.

Mineralogy, Petrology and Metamorphism

The Wissahickon Schist

Mineralogy of the schist.—The chief minerals in the Wissahickon schist are quartz, plagioclase, and micas.

PLAGIOCLASE: Throughout the schist the plagioclase is dominantly oligoclase, though the compositional range is from albite to andesine. Most of the albite occurs as porphyroblasts in the fault zone. In the garnet zone and in the fault zone, some plagioclase grains have cores

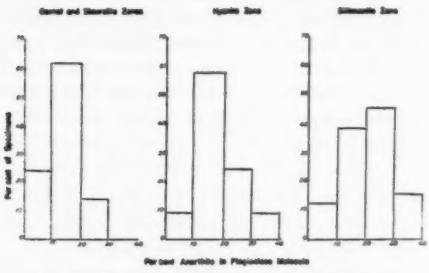

FIGURE 1.—PLAGIOCLASE IN THE SCHIST

that are slightly more calcic than the rims. This zoning and the formation of the albite porphyroblasts could be due to the introduction of soda-rich solutions during faulting or to a reconstitution of the plagioclase molecule in accordance with lower pressure-temperature conditions, attendant during faulting, than those prevailing during regional metamorphism. In a sense this process is one of retrograde metamorphism.

However, in the zones of higher metamorphic grade there is much reverse zoning of the plagioclase grains. Grubenmann (Grubenmann and Niggli, 1924, p. 425) points out that the chemical and optical differences between the more sodic core and the more calcic rim are not as great as in the normal zoning of plagioclase in igneous rocks. His explanation of the reverse zoning is that when the plagioclase grains recrystallize during metamorphism the less soluble albite forms the core of the grain and the rim is enriched in the calcic molecule. In many grains this zoning is very marked, and with the nicols uncrossed the calcic rims, having a higher refractive index, look like little raised rings around the lower middle portion.

There is a gradual increase in the anorthite content of the plagioclase with increase in metamorphic grade (Fig. 1). Recrystallization during regional metamorphism was accompanied by continual readjustment of the plagioclase molecule to the prevailing temperature and pressure conditions. In the garnet and staurolite zones, the plagioclase is albite or sodic plagioclase. In the kyanite zone, albite is less abundant and sodic oligoclase is dominant. In the sillimanite zone, calcic oligoclase is more abundant than sodic oligoclase.

The occurrence of more calcic plagioclase in the higher metamorphic grades is similar to that noted by Phillips (1930) in the Green Bed Group of the Scottish Dalradian.

BIOTITE: There is considerable variation in the pleochroism of the biotites throughout the schist. Three main types may be recognized:

a. X = yellow orange, Y = Z = red-brown.
b. X = light yellow, Y = Z = orange-brown.
c. X = greenish-yellow, Y = Z = olive-green or brown.

The pleochroism of 150 biotite specimens from the entire region was noted. The refractive index (γ) of 23 of these was determined (Table 1). In the garnet zone, the biotite is olive green or brown. Except for a few places in the fault zone, olive-green or brown is the dominant color of the biotite in the staurolite zone. Olive-green and brown biotite persists into higher zones, even into granitized areas. In the kyanite zone, orange-brown biotite is more plentiful; in the sillimanite zone, orange-brown biotite is just as abundant as olive-green and brown. The red-brown biotite was noted only in and near granitized areas, and even there is developed only locally.

Hence, as Tilley (1925) and Phillips (1930) observed, olive or greenish biotites may indicate lower metamorphic grades, and orange tints may develop as intensity of regional metamorphism increases. According to Hall (1941a), greenish and brownish tints, in general, indicate high iron content, whereas the more reddish tints are due to titania. However, no correlation can be made between the chemical composition and refractive indices because (Hall, 1941b, p. 40) ". . . a biotite which is high in iron and low in titania may have the same refractive index as one which is low in iron and high in titania". Slight variations in the proportions of iron and titania could be responsible for the similar range in refractive indices in all three colors of biotite, though the exact character

and extent of chemical variation in the biotites of the region cannot be known without chemical analyses. The increase in orange color in the zones of higher metamorphic grade may mean quartz, plagioclase, micas, and magnetite. In some places along the fault zone very large garnets, as much as 1 cm. in diameter, have developed and a few of them are considerably

TABLE 1.—*Variation of biotite in the schist*

	Olive-green and Brown		Orange-brown		Red-brown	
	Specimen	γ	Specimen	γ	Specimen	γ
Garnet Zone	*47-P495	1.620				
	52-42	1.629				
	*52-39	1.633				
	*52-100	1.638				
Staurolite Zone	52-41	1.635	*52-261	1.628		
	52-38	1.644	52-4	1.643		
Kyanite Zone	*52-53	1.629	*52-254	1.632		
	*52-54	1.629	52-28	1.638		
	52-56	1.638				
Sillimanite Zone	52-375	1.641	52-181	1.624	*47-P643	1.629
	52-21	1.643	52-394	1.628	*52-121	1.638
	52-366	1.644	52-316	1.629		
			52-22	1.632		
			52-294	1.643		

Index determined by immersion method with white light.
*Specimens have accompanying thin sections. See Plate 1 for location.

a slight decrease in iron and an increase in titania.

MUSCOVITE: Muscovite appears to be stable throughout the area, and its indices of refraction are the same for all metamorphic zones: $n_\beta = 1.605$, $n_\gamma = 1.609$. 2V is negative and about 30 degrees. The optical properties are essentially the same as those of muscovite from Dutchess County (Barth, 1936) and from the Mt. Washington Area (Billings, 1941). Muscovite commonly is intergrown with biotite. In some places in the sillimanite zone, muscovite is penetrated by needles of sillimanite or intergrown with masses or sheaves of sillimanite.

METAMORPHIC INDEX MINERALS:

1. Garnet. Large porphyroblasts of garnet, from 0.5 to 5.0 mm. in diameter, are euhedral to subhedral and contain numerous inclusions of chloritized. In thin section, the garnets are pink to colorless. They contain inclusions of quartz, feldspar, mica, and magnetite, oriented in sub-parallel contorted lines indicating a helicitic structure (Fig. 2). The parallel arrangement of the inclusions may result from an earlier schistosity in the rocks, but the S-shaped or curved pattern of the inclusions is probably the result of partial rotation of the porphyroblast during growth.

The index of refraction of garnet from all metamorphic zones was tested and found to be remarkably constant: $n = 1.805$, similar to the value of $n = 1.810 \pm .005$ for the garnets from Dutchess County (Barth, 1936, p. 786) and the garnet with an average $n = 1.803$ from the Mt. Washington Area (Billings, 1941, p. 893).

A chemical analysis* was made of garnet from Rose Glen, on the west side of the Schuylkill River, 3 miles northeast of Bryn Mawr.

	Rose Glen*		Barth (1936, p. 786)
	Analysis	Recalculated to 100%	Clove Quad., analysis
SiO_2	38.47	38.09	37.94
Al_2O_3	21.27	21.06	22.87
Fe_2O_3	6.87	6.80	—
FeO	27.07	26.80	32.45
MgO	1.70	1.68	1.83
CaO	3.97	3.93	2.04
TiO_2	0.84	0.83	1.04
MnO	0.81	0.81	1.77
	101.00	100.00	99.94

*Edith Woolever Analyst, Department of Chemistry, Bryn Mawr College.

It was difficult to get a pure sample of the garnet for analysis; even the smallest of crystals are crowded with inclusions. The analysed garnets were crushed, and as much of the quartz, biotite, and feldspar as possible was removed by bromoform separation. Nevertheless, the analysis shows an excess of SiO_2, Fe_2O_3, and MnO for the amounts needed for the garnet molecules. Recalculation of the analysis give the following:

Andradite	2.60%
Grossularite	7.87
Spessartite	1.11
Almandine	68.13
Hypersthene	4.77
Hematite	6.61
Quartz	8.91
	100.00

Considering the garnet molecules alone, the mineral is composed of:

Andradite	3.2%
Grossularite	9.9
Spessartite	1.4
Almandine	85.5
	100.00

This analysis agrees with the value for the index of refraction, according to Winchell's diagram (1933, vol. II, p. 179), and it is similar to the chemical analysis made by Barth of the garnet from Clove Quadrangle. Barth's analysis shows the almandine molecule to be about 80 per cent of the garnet.

An attempt was made to measure the specific gravity of the analyzed garnet. As the chemical analysis indicates, it was difficult to separate the inclusions completely, even from the crushed material. The chemical analysis shows about 20 per cent of material other than pure garnet. Hence, small euhedral crystals were used for specific gravity determination, and the value of 4.09 obtained was recalculated to exclude the foreign material. Assuming from inspection of thin sections (Fig. 2) that the inclusions average about 35 per cent of the crystals and assigning to them the relative proportions indicated by the chemical analysis, the recalculated specific gravity is 4.20. This probably is close to the true value. It is near the value of S.G. = 4.23 obtained by Billings (1941, p. 893) for garnet from the Mt. Washington Area.

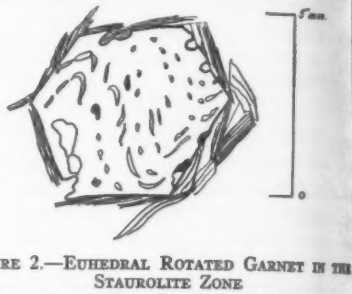

FIGURE 2.—EUHEDRAL ROTATED GARNET IN THE STAUROLITE ZONE

Thin section 52-3 (Table 6), E. side Wissahickon Creek, Chestnut Hill. The reversed S-shaped pattern of the inclusions shows clockwise rotation of the garnet porphyroblast. The sub-parallel orientation of the inclusions reveals a helicitic structure. Light inclusions are quartz and plagioclase, dark inclusions are magnetite, large grains surrounding garnet are muscovite, shaded grains are biotite.

From the above data, the garnet of regional metamorphism seems to have a rather unique and stable character throughout the metamorphic grades in which it can occur.

2. Staurolite. In many places, large dark-brown crystals of staurolite up to 2 inches long are found. Cruciform twins are common. Under the microscope, the staurolite is pleochroic, X = pale yellow, Y = golden-yellow, and Z = orange-yellow. Positive 2V is almost 90 degrees. The indices of refraction, determined on several specimens from the staurolite and kyanite zones, were constant: $n_\alpha = 1.736$, $n_\beta = 1.741$, and $n_\gamma = 1.746$. The staurolite has numerous inclusions of quartz, plagioclase, mica, and magnetite.

3. Kyanite. Blades of kyanite, as large as

0.5 cm. wide and 2 cm. long, are visible megascopically in many places. The flat crystals, light blue to deep cobalt blue, are unmistakable in the field. Under the microscope, kyanite is colorless. Negative 2V is large, and $n_\beta = 1.720$. Twins, with 001 as the twin plane, are common.

4. Sillimanite. Sillimanite occurs either as fine needles or felt-like masses intergrown with mica, or as large masses of grayish-white or pale-yellow fibrous crystals, as much as 10 cm. long and 3 cm. thick. Some large masses of sillimanite are bladed, psuedomorphous after kyanite. In thin section, fine needles of sillimanite are intergrown with mica and penetrate quartz. In some instances, sheafs or bundles of sillimanite are bent around quartz and feldspar aggregates.

OTHER ACCESSORIES:

1. Magnetite, Apatite, Zircon, and Sphene. Magnetite is in subhedral to euhedral grains, and also in minute flakes or scales, oriented along the edges or cleavages of mica. Some of the black opaque scales may be ilmenite. No test for TiO_2 was made. In the fault zone magnetite has been crushed and sheared along the mica cleavages.

Apatite is almost invariably present but always in minor proportions. The grains are colorless, very much rounded, and commonly small, from 0.01 to 0.1 cm. long.

Colorless zircon is never greater than 0.05 mm. in diameter and generally is rounded. It occurs as inclusions in the biotite, and is surrounded by strongly pleochroic haloes.

Sphene is a rare accessory, occurring in small rounded or euhedral grains, never more than 0.1 mm. long.

2. Zoisite and Epidote. Zoisite, together with some colorless clinozoisite, is an accessory more common in the schist of lower metamorphic grade. In the garnet and staurolite zones, zoisite occurs as minute, rounded, pink inclusions in quartz and plagioclase. The zoisite may form from the CaO which is in "excess of that needed to form plagioclase of a particular composition demanded by reigning physical conditions" (Turner, 1948, p. 81). This origin for the zoisite is further indicated by the fact that it is rarer in the kyanite zone, where the plagioclase has become more calcic, and in the sillimanite zone, with one exception, it is not present at all.

The exceptional occurrence of zoisite in the sillimanite zone is at Glendale Quarry, on Darby Creek, near location P 237 (Pl. 1). In this quarry a felsic layer, about 6 inches thick and 10 feet long, was observed to be rich in pink zoisite or thulite. This same layer contained pale-green crystals of epidote, 0.5–1 mm. in diameter, and green grossularite crystals, 1–3 mm. in diameter. This is the only noted occurrence of grossularite in the schist. The felsic layer may possibly represent a somewhat more calcareous bed in the original sediment.

Epidote, less abundant than zoisite and concentrated in bands parallel to the schistosity of the rocks, is confined, in general, to the sillimanite zone. In thin-sections, colorless to pale-green epidote is rounded or euhedral and shows marked cleavage in the long direction of the crystal.

3. Chlorite and Stilbite. Along the southern part of Tacony Creek (location 52–214), on a joint surface, a coating of rosettes of deep blue-green chlorite was found. The rosettes are ½–2 cm. in diameter and 1 cm. thick. The chlorite is prochlorite, and has a positive 2V of about 5 degrees. $n_\beta = n_\gamma = 1.618$. Aside from this one occurrence, chlorite was not observed megascopically in any of the schist.

In thin sections from a few locally hydrothermally altered rocks chlorite can be seen to have replaced biotite to a minor degree. In some places the chlorite is concentrated in the noses or troughs of polygonal arches or minute folds of micas. No index determination was made of the chlorite, but its pale- to bright-green pleochroism, its negative small 2V (0 to 3 degrees), and its very low birefringence and anomolous blue interference colors indicate that it is penninite.

In several places yellowish to pinkish stilbite was found in radiating masses that form coatings on joint surfaces normal to the schistosity.

4. Tourmaline. The occurrence of tourmaline seems to bear no relationship to metamorphic rank or original composition of the schist. It has been found in all metamorphic zones, though it is somewhat more prevalent as an accessory along the fault zone. Long slender black prisms, many from 0.1 to 0.5 cm. thick and 1–5 cm. long, are oriented both parallel to and across the schistosity of the rocks. In thin section tourmaline is pleochroic, with X =

pale violet gray or pink, and Z = greenish blue or dark green. The larger the grain, the deeper the colors, and crystals 1.0 mm. long are pleochroic, with X = reddish brown, and Z = black. The random orientation of the crystals and the fact that the euhedral grains are uncrushed, even in the highly sheared rocks of the fault zone, suggest that tourmaline was a late addition to the schist (Pl. 2, fig. 1).

General composition and origin of the schist.—The area of this study is one of gently rolling topography developed on the deeply weathered schist. Total relief is about 400 feet. Exposures are not abundant except along the major streams (Bascom, 1909a, p. 2–3).

The schist shows, for the most part, the results of regional metamorphism, uncomplicated by granitization or large-scale hydrothermal alterations. There are some small bodies of granodiorite in the area, and some of the hornblende gneiss in the southern and southeastern parts shows evidences of granitization. But, with the exception of a few exposures near the Schuylkill River and City Line, there are no migmatites and intruding granodiorite characteristic of the West Laurel Hill and southwestern areas (Postel, 1940).

It generally has been believed that the Wissahickon schist is of sedimentary origin, despite the fact that metamorphism has effectively removed all traces of any original sedimentary texture. The present work supports this older idea with additional data from field, microscopic, and chemical work. Moreover, the presence of the metamorphic index minerals, garnet, staurolite, kyanite, and sillimanite, in definite zonal arrangement, indicates that the regional metamorphism of these originally sedimentary rocks increased in intensity from northwest to southeast.

The rocks of the Wissahickon schist are extremely heterogeneous. Strongly schistose rocks rich in mica, alternate with more granulose rocks composed chiefly of quartz and feldspar (Pl. 2, fig. 4, and Pl. 3, fig. 2). In the field the most striking features in the greater number of outcrops are the layers of dissimilar composition, which vary in thickness from about an inch to several feet. These layers probably represent original sedimentary beds, but be

PLATE 2.—TEXTURES AND STRUCTURES IN WISSAHICKON SCHIST

FIGURE 1.—EUHEDRAL TOURMALINE IN AUGEN SCHIST

Thin section 52-90 (Table 8), Neshaminy Falls. (crossed nicols) The rock is from the fault zone. Tourmaline strongly euhedral and unbroken, although the minerals of the host rock are highly sheared and crushed. M = muscovite, P = plagioclase T = tourmaline, black grains are magnetite. Fine-grained groundmass is chiefly quartz and muscovite. Porphyroblasts of mica and plagioclase are lenticular and sheared.

FIGURE 2.—SMALL ANGULAR DISCORDANCE BETWEEN COMPOSITIONAL BANDS AND SCHISTOSITY

Thin section 52–100 (Table 5), Bryn Mawr College campus, Bryn Mawr, Pa. Note alignment parallel to schistosity of the small, irregular quartz grains in the felsic band. Micas, parallel to the schistosity, penetrate the felsic band. Q = quartz, P = plagioclase, M = micas. Solid line is parallel to compositional banding. Dashed line is parallel to schistosity.

FIGURE 3.—INCIPIENT TRANSPOSITION CLEAVAGE

Thin section 47–P268 (Table 7), ¾ mile S. of Glenside. Note that alternating bands of quartz and mica are tightly folded. There has been some breaking and slipping parallel to the axial planes of these folds, making the new cleavage. Q = quartz, M = micas, P = plagioclase, black grains are magnetite. Solid line is roughly parallel to compositional banding. Dashed line is parallel to faint new cleavage.

FIGURE 4.—COARSE-GRAINED SCHIST IN THE SILLIMANITE ZONE

Elevation looking W. Automobile keys give scale. Abandoned quarry, Logan Street and Belfield Avenue, 4 miles NW. of Frankford. Felsic layers, rich in quartz and feldspar, have been folded, broken and separated, to form light lenticular masses, around which the more plastic mica-rich layers wrap and bend.

FIGURE 5.—POLYGONAL ARCH OF MICAS IN SILLIMANITE ZONE

Thin section 52–194 (Table 8), ¼ mile N. of Tacony Creek and 2½ miles SW. of Verreeville. Note that polygonal arch, outlined by the mica laths, is less compressed than polygonal arches of micas in zones of lower metamorphic grade. (*Compare* with Figure 4.)

FIGURE 6.—SPECIMEN 52–268, R.R. CUT, ¼ MILE S.E. OF GLENSIDE

White lenticular grains are albite porphyroblasts. Dark bands are more micaceous. Light bands are more quartzose. There are fractures (light lines) along which there has been some recrystallization, creating a new schistosity. (*See also* Pl. 5, fig. 2.)

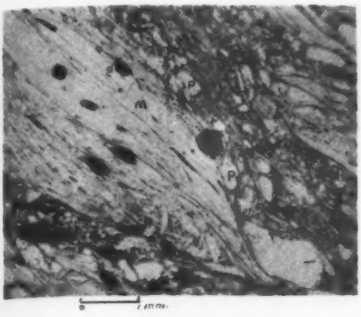

FIGURE 1.

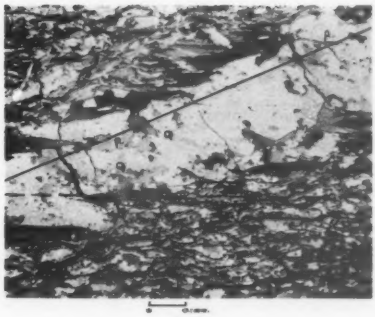

FIGURE 2.

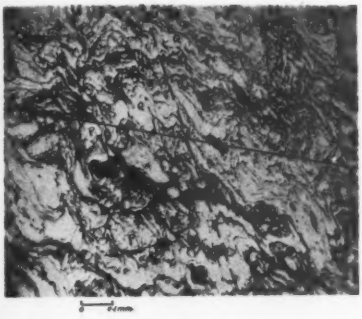

FIGURE 3.

FIGURE 4.

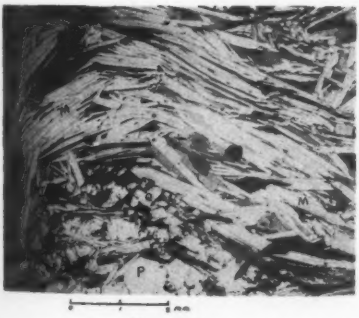

FIGURE 5.

FIGURE 6.

TEXTURES AND STRUCTURES IN WISSAHICKON SCHIST

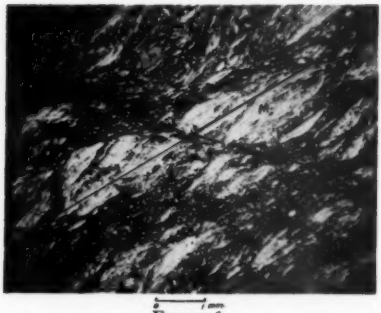

FIGURE 1.

FIGURE 2.

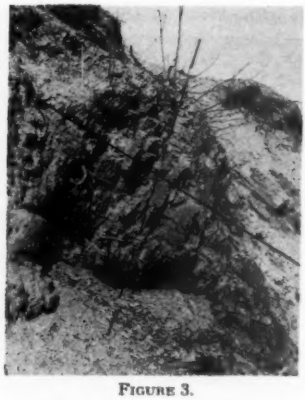

FIGURE 3.

FIGURE 4.

FIGURE 5.

FIGURE 6.

STRUCTURES IN WISSAHICKON SCHIST

ause of their intense deformation the beds cannot be treated as stratigraphic units. Within the yers there are thinner segregations or bands at are the result of metamorphic differentiion, and that have characteristic metamorphic extures and microstructures. The dominant histosity of the rocks is parallel to the bedding.

In Tables 5, 6, 7, and 8, the modes of the hist from the various metamorphic zones are ven. These modes, recalculated so that the ain constituents, quartz, feldspar, and micas, tal 100 per cent, have been plotted on the rnary diagram of Figure 3. The lack of prerred grouping of the points in any one area f the diagram is taken to indicate the irregularities of original sedimentation which graded ther completely from quartzose to arkosic to rgillaceous types. The variations in composion are in no way related to the metamorphic nes described below. Except for the domantly quartzose rocks of the garnet zone, cks that are quartz rich or mica rich or feldar rich occur throughout the region independnt of their grain-size or metamorphic rank. uartz-rich and feldspar-rich rocks will be rerred to as felsic layers. Mica-rich rocks will e called micaceous layers. In the field, felsic yers may be seen to alternate with micaceous layers, and these two rock types differ both texturally and structurally. Since the composition of the felsic and micaceous layers varies from place to place, the terms felsic or micaceous can be used only in a relative sense.

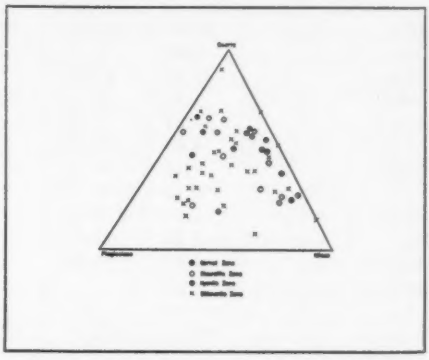

FIGURE 3.—MODES OF THE WISSAHICKON SCHIST

The chemical analyses in Table 2 are comparable to those of arkoses, impure sandstones, and shales. In Table 3 a composite analysis of 51 Paleozoic shales, an analysis of a graywacke, and a composite analysis of 371 sandstones, show strong resemblance to the schist analyses 47-P268 (from a mica-rich layer), 45-

PLATE 3.—STRUCTURES IN WISSAHICKON SCHIST

FIGURE 1.—FRACTURE CLEAVAGE IN FAULT ZONE

Thin section 52-90 (Table 8), Neshaminy Falls. Muscovite and plagioclase porphyroblasts in groundmass muscovite and quartz. M = muscovite porphyroblasts, probably sheared after growth, P = plagioclase, = tourmaline. Black grains are magnetite. Solid line is parallel to schistosity. Dashed line is parallel to acture cleavage.

FIGURE 2.—SMALL RECUMBANT FOLDS IN THE SILLIMANITE ZONE

Abandoned quarry, Logan Street and Belfield Avenue, 4 miles N.W. of Frankford. Elevation looking W. utomobile keys give scale. Light layers are feldspar-rich. Darker layers are mica-rich. Black line is parallel axial planes of folds.

FIGURE 3.—SCHISTOSITY CROSS-CUTTING LITHOLOGY IN FAULT ZONE

R.R. cut, ½ mile S.E. of Glenside. (elevation looking N.E.) Solid line is parallel to compositional banding hich strikes N. 75° E. and dips 40° SE. Dashed line is parallel to cross-cutting schistosity which strikes 75° E. and dips 75° NW.

FIGURE 4.—PARTIAL BOUDINAGE OF COMPETENT LAYERS

E. side of Old York Road, 2 miles S. of Noble. Elevation looking E. Boudins are light, lenticular masses of lsic material. Dark band is hornblende gneiss.

FIGURE 5.—RIPPLED SURFACE ON SCHISTOSITY PLANES, STAUROLITE ZONE

Dove Lake, 1½ miles N.E. of Bryn Mawr. Elevation looking E., at schistosity surfaces.

FIGURE 6.—ROTATED HORNBLENDE GNEISS BOUDIN

Hutchinson and Nedro Streets, 3½ miles N.W. of Frankford. Sillimanite zone. Elevation looking W. The -shaped pattern (shown by black line) of the foliation in the hornblende gneiss indicates a clockwise rotaon of the boudin. In the surrounding schist there has been an overthrusting to the south and an underhrusting to the north. This is consistent with the general movement shown by the overturning of the folds the southeast in the sillimanite zone.

TABLE 2.—*Chemical analyses and modes of specimens in the Wissahickon Schist*

	Garnet zone	Staurolite zone		Kyanite zone		Sillimanite zone	
	52-100	44-42	44-48a	44-9	47-P268	52-15	45-T26
Chemical Analyses							
SiO_2	78.48	67.40	81.90	69.85	60.33	69.94	73.68
TiO_2	0.61	0.75	0.60	1.46	1.41	0.75	0.81
Al_2O_3	11.56	16.61	8.59	13.74	20.85	14.12	12.49
Fe_2O_3	2.57	3.76	0.47	1.82	3.59	1.28	2.10
FeO	0.69	3.58	2.10	3.08	4.47	4.10	2.22
MnO	0.03	0.11	0.03	0.08	—	0.08	0.18
MgO	0.12	1.36	0.13	2.34	2.07	1.61	2.04
CaO	0.40	0.58	1.67	0.78	1.82	1.29	0.56
Na_2O	1.41	0.87	2.98	2.34	1.38	2.31	2.97
K_2O	2.26	3.25	1.16	2.57	2.87	3.66	2.91
H_2O	1.97	1.84	0.62	1.36	2.78	1.46	1.34
P_2O_5	—	—	—	0.11	0.28	—	0.12
	100.10	100.11	100.25	99.53	101.85	100.60	101.42
Modes							
Quartz	64	56	57	49	37	35	45
Plagioclase	3 (An_7)	5 (An_{34})	38 (An_6)	20 (An_{14})	4 (An_9)	37 (An_{34})	6 (An_{2}
Biotite	7	4	2	12	11	19	18
Muscovite	24	22	—	18	46	8	28
Accessories	2	13	3	1	2	1	3

52-100. Analyst, Phyllis Vipond, Dept. of Chemistry, Bryn Mawr College. Bryn Mawr College Campus 20′ west of Park Hall.

44-42. Analyst, Ruth Stoddard, Bryn Mawr College (Cloos and Heitanen 1941) Dove Lake, 2 mi N. E. of Bryn Mawr.

44-48a. Analyst, Elizabeth M. Hardy, Bryn Mawr College (Cloos and Heitanen, 1941): ½ mile west Rose Glen.

44-9. Analyst, Olive Bates, Dept. Chemistry, Bryn Mawr College (Cloos and Heitanen 1941), Foxcroft

47-P268. Analyst, F. A. Genth, Florence Bascom's notes, W. side Tacony Creek, 1 mile S. E. of Glensid

52-15. Analyst, Peggy Quin, Dept. Chemistry, Bryn Mawr College, Wissahickon Creek, ½ mi. N. E. City Line and Schuylkill River.

45-T26. Analyst, F. A. Genth, Jr., (Bascom, 1905, p. 304) Neshaminy Creek, ½ mi. south of Neshami Falls.

T26 (from a mica-rich layer), and 52-100 (from a quartz-rich layer) respectively. Recrystallization of such sediments as those in Table 3 could produce the schists in Table 2 without significant addition or subtraction of material. In the adjacent region of granitized schist, where potash feldspar is an important constituent, some alteration of the original composition of the sediments may have occurred, but since no chemical analyses have been made of the migmatites the actual amount and character of the chemical change is not known.

Metamorphic zones.—The presence of the metamorphic index minerals, garnet, staurolite kyanite, and sillimanite, in definite zonal a rangement, indicates a regional metamorphism which increased in intensity from northwest t southeast. From about 250 plotted points, it possible to draw staurolite, kyanite, and sill manite isograds across the area. Petrologic an mineralogic changes from zone to zone ar further evidences of progressive regional meta morphism.

For the most part, the metamorphism i "isochemical" (Barth, *et al.*, 1939, p. 339). Th rocks in the garnet zone belong to the epidot

TABLE 3.—*Chemical analyses of sediments*

	A, 51 Paleozoic Shales	H, Graywacke	G, 371 Sandstones
SiO_2	60.15	76.84	84.86
TiO_2	0.76	—	0.41
Al_2O_3	16.45	11.76	5.96
Fe_2O_3	4.04	0.55	1.39
FeO	2.90	2.88	0.84
MnO	tr.	tr.	tr.
MgO	2.32	1.39	0.52
CaO	1.41	0.70	1.05
Na_2O	1.01	2.57	0.76
K_2O	3.60	1.62	1.16
H_2O	4.71	1.87	1.74
P_2O_5	0.15	—	0.06
CO_2	1.46	—	1.01
SO_3	0.58	—	0.09
LiO_2	tr.	—	tr.
BaO	0.04	—	0.01
Cl	—	—	tr.
C	0.88	—	—
	100.46 (Compare with 47-P268)	100.18 (Compare with 45-T26)	99.86 (Compare with 52-100)

A. Clarke (1924, p. 552).
H. Clarke (1924, p. 547).
G. Clarke (1924, p. 547).

amphibolite facies of Eskola (Barth, *et al.*, 1939, p. 360), and correspond to the chloritoid-almandine sub-facies of Turner (1948, p. 89–90). The rocks of the staurolite, kyanite, and sillimanite zones belong to the amphibolite facies as defined by Eskola (Barth, *et al.*, 1939, p. 351–355) and correspond to the staurolite-kyanite subfacies and the sillimanite-almandine sub-facies of Turner (1948, p. 81–87). According to Grubenmann's (Barth, *et al.*, 1939, p. 375) concept of depth zones, the meso-zone (in part) and the kata-zone (in part) are represented (Table 4). Most probably the recrystallization of the sediments took place at considerable depths and under conditions of high temperature and pressure.

GARNET ZONE: The garnet zone occupies a small portion of the area. It is three-quarters of a mile wide, and 6 miles long, and lies just south of the Rosemont fault. This zone represents the lowest grade of metamorphism in the area.

Near the fault the rocks of the garnet zone have a remarkably uniform appearance, which may in part be due to the fault movement and in part to original compositional uniformity (Table 5). The rocks are all fine-grained, very schistose, and look platy or like a pack of cards which have been pushed past each other. This appearance is due to the concentration, through mechanical differentiation during deformation, of the micas in numerous parallel thin bands. The arrangement of the micas gives the rocks their marked schistosity but their chief constituent is quartz, which also is concentrated in bands. At first glance, thicker bands look like quartz veins, but they are discontinuous and, like the thinner quartz bands, probably are segregations of an originally more uniform rock.

Garnet porphyroblasts are visible in the hand-specimen.

The microstructure of the rocks in the garnet

TABLE 4.—*Metamorphic zones and mineral facies*

Eskola (Barth, *et al.* 1939, p. 360)		Turner (1948, p. 76–90)	Grubenmann (Grubenmann and Niggli, 1924, p. 398–399)	Corresponding metamorphic zones in the Wissahickon schist at Philadelphia
Facies	Zone	Facies	Depth zones	Zones
Epidote-amphibolite facies	Almandine zone	*Albite-epidote amphibolite facies* Chloritoid-almandine sub-facies	Meso-zone (in part)	Garnet zone
Amphibolite facies	Hornblende zone	*Amphibolite facies* Staurolite-kyanite sub-facies		Staurolite zone Kyanite zone
		Sillimanite-almandine sub-facies	Kata-zone (in part)	Sillimanite zone

zone is schistose, with a granoblastic texture in the felsic bands. Quartz is the chief constituent. Some plagioclase shows faint zoning with the centers of the grains a little more calcic. Muscovite and biotite are concentrated in thin parallel bands. Biotite and muscovite generally are intergrown or in parallel position, suggesting simultaneous recrystallization. Some muscovite, however, cross-cuts the schistosity, creating the appearance of a faint fracture cleavage. The orientation of these cross-cutting grains may be due to their recrystallization during movement later than that which produced the main schistosity, or to the rotation of the mica during continuing movement, or it may be a mimetic reproduction of an earlier cross-cutting schistosity. This is suggested by pre-crystalline deformation of some of the mica (Fig. 4). A few thin sections show some tightly-compressed microfolds, with the micas recrystallized into polygonal arches. The microfolds, possibly, are the remnants of a previous structure that was destroyed by the intense shearing movement connected with the faulting nearby. In general, the schistosity is parallel to the compositional layers, but in one thin section (Pl. 2, fig. 2) there is an angular discordance of about 10 degrees between the schistosity and the parallel felsic and micaceous layers.

TABLE 5.—*Modes of schist in the garnet zone*

	Number of specimen				
	52-39	47-P1	52-100	44-52	47-P93
Quartz	67	66	64	58	57
Plagioclase	8	11	3	6	4
Biotite	10	14	7	9	21
Muscovite	12	7	24	26	9
Accessories*	3	2	2	1	9
Magnetite	X	X	X	X	X
Zircon	X	X	X	X	
Garnet	X	X			X
Apatite		X	X	X	X
Sphene			X		
Chlorite	X	X		X	X
Zoisite				X	
Sericite				X	
Tourmaline					X
	% An in plagioclase				
	17	15	7	20	6
	Grain size in mm.				
Quartz, diameter	0.05	0.1–0.3	0.02–0.1	0.1–0.3	0.05–0.1
Plagioclase, diameter	0.1	0.1–0.3	0.05–0.1	0.1–0.3	0.05–0.1
Mica, length	0.2	0.3	0.1–0.3	0.2–1.5	0.1–0.3
Garnet, diameter	0.5	0.2			1.0–2

* Presence of individual accessories marked by X.

Magnetite, zircon, and apatite are common accessories. Sphene is rare. Tourmaline occurs in some rocks. Pink zoisite or thulite, in minute rounded grains, is scattered throughout the quartz and feldspar. Minor chlorite is an alteration of biotite.

Euhedral garnet contains numerous inclusions of quartz, feldspar, magnetite, and some mica, in sub-parallel lines which make an angle with the schistosity of the host rock. Their orientation suggests a helicitic structure.

STAUROLITE ZONE: The staurolite zone covers an area 1½–2 miles wide and 14 miles long. The descriptions below pertain to all rocks in this area except those directly along the Cream Valley fault.

In the staurolite zone, the rocks are strongly schistose. They are megascopically coarse grained and do not have as uniform an appearance as the rocks in the garnet zone. The variation in sedimentation, graphically represented in Figure 3, (*see also* Table 6) is quite obvious in the alternating felsic and micaceous layers. Within these layers, thinner bands represent mechanical differentiates produced during deformation. Numerous plications or wrinkles occur in the micaceous layers, which, under stress, are less competent than the felsic layers.

Garnet locally has attained its greatest development in the staurolite zone. In many places staurolite also is visible megascopically.

The microstructures in the staurolite zones are schistose in the micaceous layers and granulose in the felsic layers. The texture in the felsic layers is granoblastic. The megascopically coarse grains of quartz, feldspar, and mica under the microscope are revealed to be aggregates of smaller grains. The irregular boundaries of quartz grains are not as markedly sutured as in the garnet zone. Plagioclase is chiefly oligoclase and shows some "reverse" zoning (Phillips, 1930, p. 249). The micas have recrystallized into polygonal arches that are less compressed than those in the garnet zone. In some places micas on the convex side of the polygonal arches are bent or show wavy extinction.

The accessories in the staurolite zone are the same as in the garnet zone. There is also some rare epidote.

KYANITE ZONE: The kyanite zone is from ½–2 miles wide and 15 miles long.

The structure is dominantly schistose in the kyanite zone; as in the staurolite zone, alternating felsic and micaceous layers represent original compositional differences (Fig. 3; Table 7). The rocks are coarse-grained, especially in the micaceous layers, where plates of mica seem to merge one into the other, forming large, silvery, crinkled or rippled sheets between the felsic layers. Crystals of garnet and staurolite, as well as blades of kyanite, are visible megascopically in many places. Where these minerals are abundant, the surfaces of schistosity have a rough or hackly appearance.

In the kyanite zone, the microstructure is schistose in the micaceous layers and gneissose

FIGURE 4.—POLYGONAL ARCH OF MICAS IN GARNET ZONE

Thin section 44–52 (Table 5), W. side of Schuylkill River, at Lafayette, Penna. Large light grains are muscovite, shaded grains are biotite, irregular grains are quartz and plagioclase. Compare with Plate 3, figure 5, showing more open polygonal arch.

or somewhat cataclastic in the felsic layers. Texture in most of the felsic layers is granoblastic. The boundaries of quartz grains are not as irregular as those in the garnet and staurolite zones. Plagioclase, generally sodic to calcic oligoclase, shows reverse zoning. In some of the felsic layers, quartz and plagioclase grains appear crushed. In the micaceous layers, the micas are arranged in open polygonal arches or in sub-parallel bands that are wavy or folded. Much of the mica has recrystallized into straight grains. However, in the folds and convex sides of the polygonal arches, many mica grains are curved and show undulating extinction.

In one thin section (47-P268) from a mica-rich layer, an interesting microstructure, resembling transposition cleavage, was noted. The micaceous and quartzose bands are tightly folded. The micas have been bent, broken, and then pulled apart, with some recrystallization taking place along the direction of fracture (Pl. 2, fig. 3). The quartz, in minute crushed grains, also appears to have begun recrystallization along the new cleavage that crosscuts

TABLE 6.—*Modes of schist in the staurolite zone*

	Number of specimen											
	52-261*	52-52	44-48a	52-47	52-252	44-42	52-1*	52-8*	52-245	52-268*	52-26	52-3
Quartz	63	63	57	57	56	56	45	41	28	26	21	21
Plagioclase	22	18	38	24	10	5	26	6	16	11	50	8
Biotite	3	5	2	8	11	4	10	13	16	42	6	21
Muscovite	6	12	—	10	22	22	16	32	34	21	22	39
Accessories	6	2	3	1	1	13	3	8	6	1	1	11
Magnetite		X	X	X	X	X	X	X	X	X	X	X
Zircon	X	X		X	X	X	X	X	X	X	X	X
Garnet	X		X		X	X		X	X			X
Staurolite			X									X
Apatite	X	X	X	X	X		X	X	X		X	X
Sphene			X				X					
Chlorite	X				X	X	X		X		X	X
Zoisite			X			X						
Epidote			X									
Sericite	X		X					X				X
Tourmaline					X	X	X	X	X		X	X
	% An in plagioclase											
	22	16	6	9	17	24	17	15	23	6, 14	12	14
	Grain size in mm.											
Quartz, diameter	0.05	0.05–0.2	0.05–0.3	0.1–0.3	0.05–0.1	0.05–0.1	0.05–0.1	0.02	0.1	0.005	0.1–0.3	0.1
Plagioclase, diameter	0.05–0.3	0.05–0.2	0.05–0.3	0.1–0.3	0.1–0.2	0.2–0.3	0.1–0.2	0.05	0.1	0.02–0.2	0.1–0.5	0.2
Mica, length	0.3	0.1–0.2	0.05–0.3	0.3	0.2–0.5	0.2	0.5–1.0	0.2–0.2	0.2	0.1–0.3	0.4	0.5–1.5
Garnet, length	0.5–1.0		0.1–1.0		0.3	1.0–3		1.0	1.0			3–5

* In fault zones.

the direction of the original compositional bands. This is the only case in which a schistosity that might be called transposition cleavage has been observed in thin section.

Accessories are the same as in the zones of lower metamorphic grade. Both garnet and staurolite persist into the kyanite zone. In thin section, the garnet porphyroblasts are less strongly euhedral than in the garnet or staurolite zones. The irregular grains may result from partial resorption of the garnet, which, at the higher grade of metamorphism, had begun to be out of equilibrium.

Zoisite is rarer in the kyanite zone than in the zones of lower metamorphic grade. In one or two places, rounded or subhedral crystals of zoisite are concentrated in bands parallel to the schistosity.

SILLIMANITE ZONE: The sillimanite zone covers almost two-thirds of the area studied; in most of the zone, it does not seem possible to relate the occurrence of sillimanite to the intrusive granodiorites.

In the granitized schists to the southwest of this region, some of the sillimanite present may be due to the intrusion of granitic magma. Andalusite, a common associate of sillimanite formed by thermal metamorphism (Turner, 1948; Harker, 1933), has been found near by in Delaware County, (E. H. Watson, Personal communication). However, in the region limited by this paper no andalusite has been noted.

Moreover, sillimanite where observed as a product of thermal metamorphism is found near granitic intrusions. Watson (1948), in describing the migmatites of Kildonan, Sutherland, states that sillimanite, or fibrolite, is not present outside the injection area, and Smith (1945), in studying some schists in South Carolina, finds the sillimanite confined to the schist near

TABLE 7.—*Modes of schist in the kyanite zone*

	Number of specimen							
	52-11	52-362	44-9	52-30	47-P268	52-198	52-140*	52-29
Quartz	65	56	49	45	37	26	23	19
Plagioclase	29	28	20	40	4	1	3	40
Biotite	5	10	12	12	11	34	21	17
Muscovite	—	—	18	2	46	34	47	24
Accessories	1	6	1	1	2	5	6	<1
Magnetite	X	X	X	X	X	X	X	X
Zircon	X	X	X	X	X	X	X	X
Garnet	X				X		X	
Staurolite	X				X			
Kyanite						X		
Apatite	X	X	X		X	X	X	
Chlorite			X				X	
Zoisite			X					
Epidote		X				X		
Tourmaline					X	X	X	X
	% An in plagioclase							
	14	25-32	14	14	7	17	24	11
	Grain size in mm.							
Quartz, diameter	0.1–1.0	0.05–0.2	0.1–0.3	0.1–0.3	0.005	0.1–0.5	0.02	0.1–0.5
Plagioclase, diameter	0.1–1.0	0.05–0.2	0.1–0.3	0.1–0.3	0.1	0.1–0.5	0.1–0.3	0.1–0.5
Mica, length	0.2–0.4	0.05–0.2	0.3–0.8	0.1–0.5	0.2	1–2	0.5–1.0	0.2–1.0
Garnet, diameter	1.0				0.2		0.2	

*In fault zone.

granite, either up to 200 feet from granite contacts, or in roof pendants of underlying granite at no great depth.

However, granodiorite bodies within the area studied are small and few, and the most abundant occurrences of sillimanite are at considerable distances from the granodiorites. Also, no sillimanite has been found near the small granodiorite bodies in the staurolite zone. The origin of small knots of quartz and sillimanite (Faserkiesel, Rüger, 1931) found in great abundance on the east side of the Schuylkill River, 1½ miles southeast of City Line (Pl. 1, Location P125), may be connected with the surrounding migmatites, but most of the sillimanite throughout the region studied originates from the processes of regional metamorphism alone.

Barth (1936) uses the sillimanite isograd as a boundary line between schists and gneisses. However, in the sillimanite zone of this region, abundant micaceous layers have a schistose structure, though in felsic layers the structure is dominantly gneissose. Hence, the term schist, which is definitely applicable to the rocks of lower metamorphic grades, will be retained for the rocks of the sillimanite zone.

There is a marked coarsening of the rocks in this zone. Probably original compositional differences were emphasized by mechanical differentiation during metamorphism, and even bands of originally uniform composition may have been subjected to a "tectonic unmixing" (Knopf and Ingerson, 1938, p. 201). One of the common changes of metamorphism is the increase of grain size through recrystallization. It is highly probable that the high temperatures prevailing at the highest grade of metamorphism released the water of composition from the sediments, and that these hot solutions plus others, possibly connected with the migmatization to the southwest, further aided the process of recrystallization and metamorphic differentiation.

Recrystallized quartzose layers look very much like quartzite. Feldspar-rich layers have been broken and drawn apart during deformation to form individual masses or lenticles that give the rocks the appearance of a pegmatized schist (Pl. 2, fig. 4). However, no potash feldspar is present. Extremely coarse feldspathic and quartzose masses are common. At first glance, these masses look like pegmatites, but plagioclase and not potash feldspar is the feldspathic constituent. The very coarse-grained masses are composed of the same minerals, in approximately the same proportions, as the fine-grained rocks of lower metamorphic grade (Table 8).

Garnet is not as abundant as in rocks of lower metamorphic grades. Kyanite is visible megascopically in some rocks. Sillimanite is present in large masses in many rocks and is abundant in both felsic and micaceous layers. Harker (1932, p. 227–228) suggests two possible sources of sillimanite in regionally metamorphosed rocks: (1) kyanite which gives place to sillimanite in the highest grade of metamorphism, and (2) muscovite. As regional metamorphism reaches its highest intensity, stress begins to decline and permits the dissociation of muscovite, thus forming some potash-feldspar and releasing some free alumina. In this region, potash-feldspar is not a constituent of the schist of the sillimanite zone, but sillimanite pseudomorphs after kyanite have been found.

Microstructure in the sillimanite zone is schistose in the micaceous layers and granulose or cataclastic in the felsic layers. In thin-sections, lenticles of quartz and plagioclase, up to 1 cm. in diameter are seen to be composed of many smaller grains. Deformation, which continued throughout recrystallization, served to break down some of the more felsic constituents into smaller grains surrounding larger grains. One thin section (52–109) from a felsic layer 200 yards from the Holmesburg granodiorite, shows very marked granulation, with mosaic texture and uniform grain-size. The intrusion of the granodiorite, in this one instance, could have caused the granulation of the surrounding rock.

In some thin sections, quartz grains show wavy extinction and plagioclase grains look "patchy" or have an irregular extinction. Twin lamellae of the plagioclase frequently are broken or bent. The plagioclase, in general, is calcic oligoclase to sodic andesine, and reverse zoning is common. The micas occur in sub-parallel discontinuous lines or in gentle folds. Some of the micas have recrystallized into open polygonal arches (Pl. 2, fig. 5), but many mica grains are bent and show wavy extinction. Accessories are for the most part the same as those in the other metamorphic zones. There is no zoisite, but

TABLE 8.—*Modes of schist in the sillimanite zone*
(Arranged according to decrease in quartz content)

TABLE 8.—*Modes of schist in the sillimanite zone*
(Arranged according to decrease in quartz content)

	Number of specimen													
	52-178	52-96	52-104*	52-17	52-225	52-98	47-P541*	52-109	52-128	47-280b	52-91*	45-T26	52-170	52-130
Quartz	88	67	67	66	59	55	54	48	47	47	45	45	51	41
Plagioclase	8	17	—	26	23	18	12	—	28	13	26	6	35	23
Biotite	2	3	7	6	1	16	13	22	8	13	11	18	7	15
Muscovite	1	10	23	—	5	9	8	23	14	11	9	28	10	17
Accessories	1	3	3	2	12	2	12	7	3	16	9	3	7	4
Magnetite	X	X	X	X	X	X	X		X	X	X	X	X	X
Zircon	X	X	X	X	X	X	X		X		X	X	X	X
Garnet		X	X	X	X	X	X			X	X		X	X
Staurolite		X					X							
Kyanite														
Sillimanite							X						X	
Apatite		X	X	X		X		X	X	X	X	X	X	
Sphene														
Chlorite	X				X		X			X			X	X
Epidote								X						
Sericite		X			X				X				X	X
Tourmaline			X				X						X	X
	% An in plagioclase													
	12	12	—	22	12	14	18	—	28	12	32	12	15	18
	Grain size in mm.													
Quartz, diameter	0.05–1.0	0.05–0.2	0.05	0.05–0.2	0.05–0.5	0.05–0.2	0.05	0.05	0.02–0.1	0.02–0.05	0.05–0.1	0.05	0.05–0.2	0.1
Plagioclase, diameter	0.1–0.2	0.1–0.3	—	0.1–0.3	0.1–0.4	0.1–0.3	0.05	—	0.2–0.5	0.02–0.05	1.0	0.1	0.5–1.0	0.2–0.5
Mica, length	0.1–0.3	0.5–1.5	0.1–0.3	0.2	0.2–0.5	0.1–0.3	0.3	0.1–0.2	0.2–0.5	0.05–0.1	0.05–0.1	1.0–0.3	2.0–3.0	0.2–0.8
Garnet, diameter		0.5–1.0	0.5–1.0	0.3–0.5	1.0–2.0	0.5–1.0	0.3	—		0.5–1.0	1.5		1.5	0.5

TABLE 8—*Continued*

	Number of specimen																	
	52-132	52-136	52-88*	52-153	52-152	52-15	52-143	52-107	52-21	52-194	52-141*	47-P546	52-82	47-PBB2	47-P604	52-150	52-73*	52-90*
Quartz	39	37	37	36	36	35	29	28	28	28	28	24	23	21	18	16	8	7
Plagioclase	44	39	16	50	19	37	45	46	34	10	4	55	50	50	33	55	—	28
Biotite	7	13	18	6	21	19	25	14	13	27	—	20	25	21	38	16	13	27
Muscovite	9	10	25	6	20	8	—	5	21	33	63	—	—	2	—	13	66	35
Accessories	1	1	4	2	4	1	1	7	4	2	5	1	2	6	11	<1	13	3
Magnetite	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X
Zircon	X	X	X	X	X	X	X	X	X	X		X	X	X		X		X
Garnet	X		X	X		X	X	X	X	X			X		X		X	
Staurolite																		
Kyanite											X							
Sillimanite									X					X				
Apatite	X	X	X	X	X		X	X	X		X	X	X	X	X	X		X
Sphene																X		
Chlorite			X	X	X				X					X			X	
Epidote					X													
Siricite	X	X				X	X	X	X			X			X			X
Tourmaline	X		X		X					X	X						X	
	% An in plagioclase																	
	23	26	16	24	25	24	14	21	24	24	18	24	25	17	24	23	—	25
	Grain size in mm.																	
Quartz, diameter	0.05–0.4	0.05–0.4	0.005	0.05–0.2	0.1–0.5	0.2	0.1	0.2	0.1	0.2–0.5	0.005	0.5	0.05–0.3	0.05–0.3	0.1	0.05–0.1	0.1	0.02–0.05
Plagioclase, diameter	0.1–0.5	0.1–0.5	0.05–0.2	0.2–0.5	0.5–1.5	0.2	0.2	0.5	0.2	0.2–0.5	0.005–0.05	0.8	0.2–0.5	0.1–0.3	0.2	0.2	0.1	0.05–0.3
Mica, length	0.5–1.0	0.5–1.0	.005 & 1.0	0.5–1.0	0.5–1.5	0.2–0.8	0.5–1.0	0.5–1.5	0.3–1.0	0.5–1.0	0.05	0.5–1.0	0.1–0.8	0.2–1.0	0.2–0.5	0.2–0.5	0.2 & 1.0–3.0	0.3
Garnet, diameter	0.7		0.5–0.7	0.5		0.3	1.0	0.1	1.0	1.0			0.8		0.5–1.0		0.1–1.5	

* In fault zone.

epidote is found in a few rocks. Garnet porphyroblasts are very irregular and broken. Many of them have their long direction parallel to the schistosity. Remnants of staurolite and kyanite grains have been found in a few thin sections. Both staurolite and kyanite persist into the sillimanite zone, but, in most cases, heavy mineral separations were necessary to determine their presence.

The fault zone.—Along the Cream Valley-Huntingdon Valley fault the rocks, called the Spangled Schists by Rand (1901), have an appearance which is markedly different from that of the schist throughout the rest of the area. The zone is only 50 to at most 150 feet wide, but it extends from the Delaware River at Morrisville southwest for 27 miles to Chestnut Hill, where it meets a branch fault, the Rosemont fault (Armstrong, 1941). Both faults truncate the zones of regional metamorphism (Pl. 1, 4).

The rocks in the fault zone are intensely sheared and cross-cut by a fracture cleavage. The characteristic platy appearance, noted also in the rocks throughout the greater portion of the garnet zone, is very marked in the fault zone and dies out within 100 feet of the fault zone along the Cream Valley-Huntingdon Valley fault.

The original lithological differences seem to have had some control over the reaction of the rocks to the fault movement, and these changes are independent of metamorphic rank. In Tables 6, 7, and 8, the modes of specimens from the fault zone have been starred.

Quartz and feldspar-rich layers have become mylonized or have recrystallized into augen schists. Quartz-rich layers have been crushed to form extremely fine-grained dense-looking rocks. In the more feldspathic rocks there are many porphyroblasts of plagioclase (Pl. 2, fig. 6). The more micaceous layers have become either fine-grained, gray-green muscovite schists, or extremely coarse-grained spangled schists in which the porphyroblastic muscovite is in plates up to 1 cm. in diameter. There seem, then, to have been two distinct, though possibly partly contemporaneous, processes during faulting; a destructive process, involving reduction of grain-size and milling down of the mineral constituents, and a constructive process, in which solutions aided recrystallization of some of the minerals into porphyroblasts. The sheared and lenticular character of the porphyroblasts (Pl. 2, fig. 1) suggests that the destructive process or milling down of the constituents continued after the formation of the porphyroblasts. The occurrence of porphyroblasts of plagioclase or muscovite does not indicate necessarily the introduction of material. The solutions which aided the recrystallization could have come from the rocks themselves (Barth, 1936, p. 803–806). Water may have been released from the hydrous minerals by frictional heat generated during faulting.

The microstructure of the rocks in the fault zone is schistose or granulose. In the quartzose layers, very small quartz grains have extremely irregular or sutured boundaries, and even the smallest grains show strain shadows. Feldspathic layers have both cataclastic and porphyroblastic textures. Plagioclase lenticles, in general, appear to be the products of recrystallization rather than more resistant material which has not been completely crushed. It is true that in some thin sections plagioclase grains are fractured or have bent twin-lamellae or show wavy extinction, all of which are the results of crushing; but most of the larger plagioclase grains are unbroken and occur in a ground mass chiefly of quartz and muscovite. This seems to suggest that the bulk of the plagioclase substance in the rock went into the making of the porphyroblasts. In some places, the porphyroblasts are albite and have a poeciloblastic texture due to many inclusions of oligoclase. Some plagioclase shows faint zoning, with the rims either albite or a sodic oligoclase and the cores a medium or calcic oligoclase. This zoning, as well as the albite porphyroblasts, may be due to the process of metamorphic diffusion rather than to the addition of soda-rich solutions. The recrystallization of a more sodic plagioclase than that originally formed in the schist may be a retrograde metamorphic process connected with the faulting. The more sodic plagioclase porphyroblasts or the sodic rims around more calcic plagioclase grains are evidences of lower pressure-temperature conditions than those prevailing during the formation of the metamorphic zones.

The fine-grained greenish schists which, in the field, appear to be chlorite schists are revealed, under the microscope, to be composed chiefly of very fine scales of muscovite and some

porphyroblastic muscovite. The texture of these schists is lepidoblastic. Magnetite flakes, most probably the result of the grinding down of larger magnetite grains, are arranged like little beads on a string or in many parallel and folded lines. Narrow discontinuous bands of crushed quartz also outline faintly a pattern of folding. In one thin section a few albite porphyroblasts contain numerous inclusions of small pale-green euhedral tourmaline. Garnet porphyroblasts are strongly euhedral and contain minute magnetite inclusions arranged in parallel and folded lines. In the coarse-grained or "spangled" micaceous schists, the muscovite is almost entirely porphyroblastic. Lenticular masses of muscovite are bent into tight folds, suggesting that movement continued after their recrystallization. The schistosity, formed by the parallel arrangement of the muscovite porphyroblasts, is crosscut by a marked fracture cleavage, along which there has been some recrystallization into a new schistosity (Pl. 3, fig. 1).

A few garnets in the coarser-grained schists show alteration to chlorite. Kyanite, staurolite, and some sillimanite have been noted in a few thin sections. They are very much crushed but show no other alterations. Tourmaline is a common accessory in the fault zones. It is everywhere euhedral, even where the other constituents of the rocks are fragmented. Possibly it was introduced after movement.

Sequence of recrystallization in the schist.—Most of the recrystallization of the minerals in the schist took place during movement. Table 9 shows that in each metamorphic zone, and the fault zone as well, there was some deformation contemporaneous with recrystallation. The garnet zone has been omitted from the table, since, because of its proximity to the Rosemont fault, its original sequence of recrystallization has been masked by the recrystallization due to later fault movement.

TABLE 9.—*Relationship of recrystallization to deformation in the schist*

	Staurolite zone	Kyanite zone	Sillimanite zone	Fault zone
Quartz	Para- and post-deformation	Pre- and para-deformation	Pre- and para-deformation	Pre-deformation
Plagioclase	Para- and post-deformation	Pre- and para-deformation	Pre- and para-deformation	Pre- and para-deformation
Micas	Para- and post-deformation	Para-deformation	Pre- and para-deformation	Pre- and para-deformation
Garnet	Pre- and para-deformation	Pre- and para-deformation	Pre- and para-deformation	Pre-deformation
Staurolite	Para-deformation	Pre- and para-deformation	Pre- and para-deformation	Pre-deformation
Kyanite	—	Para-deformation	Pre- and para-deformation	Pre-deformation
Sillimanite	—	—	Para-deformation	Pre-deformation

In many places, micas wrap and bend around recrystallized aggregates of quartz and feldspar. In the staurolite zone, recrystallization of the micas appears undisturbed, but in the zones of higher metamorphic grade, bent and folded mica grains reveal continuing reaction to deformation after recrystallization had ceased. Petrofabric analysis by Hietanen (1941, p. 150–184) of rocks from several areas west of the Schuylkill River show several generations of quartz and mica recrystallization. From inspection of thin sections, the various stages of recrystallization are more obvious in the micas than in the quartz. Most of the micas are parallel to a main schistosity or S. Whether this S-plane is S_1, a bedding schistosity, or S_2, a shear cleavage, is difficult to determine without petrofabric analysis. In the field the main schistosity appears parallel to the compositional banding, but Hietanen's analyses revealed an angular discordance of 20 degrees between the compositional bands and the dominant schistosity. Under the microscope such an angular discordance was noted in only one thin section (see description of garnet zone, thin section 52-100, and Pl. 2, fig. 2). In all other thin

sections, and in the field, the dominant schistosity is parallel to the compositional layers. It is possible that, throughout most of the area, increase in intensity of deformation reduced to zero the angular discordance noted by Hietanen, and the S_1 and S_2 merged into what Turner (1948, p. 280) calls a "compromise schistosity."

A faint S_3, or cross-cutting schistosity is created by the diagnonal arrangement of some mica flakes across the S-plane. In the fault zone this is probably due to recrystallization along planes of fracture cleavage produced by shear movement during faulting. Throughout the rest of the area, the S_3-plane probably is a reflection of the rotation of some of the mica flakes into a new position during deformation, which continued throughout recrystallization.

Porphyroblasts of garnet and staurolite contain inclusions of quartz, feldspar, micas, and magnetite. Hence the recrystallization of the metamorphic index minerals followed that of the inclusions. In the higher metamorphic zones, the order of recrystallization is kyanite and then sillimanite. This is borne out by sillimanite pseudomorphs after kyanite.

There are very local hydrothermal alterations. Minor chlorite formed at the expense of biotite, and plagioclase shows sericitization along cleavage cracks. In a few places, quartz occurs in late introduced veins. Strongly euhedral tourmaline, completely unfractured, even in the crushed rocks of the fault zones, probably was the last mineral to form and may have been introduced hydrothermally or pneumatolytically after all movement had ceased.[1]

The probable sequence of recrystallization is as follows:

1. quartz and feldspar 2. biotite and muscovite	parallel to S_2
3. quartz and feldspar 4. biotite and muscovite	parallel to "compromise schistosity"
5. garnet	
6. staurolite	
7. kyanite	
8. sillimanite	
9. biotite and muscovite	parallel to S_3
10. chlorite	
11. sericite	
12. vein quartz	
13. tourmaline	

[1] In the Chickies quartzite near by, tourmaline crystals are broken or "stretched", showing that their crystallization was previous to regional deformation. The tourmaline of the Wissahickon schist is a post-deformation mineral and hence cannot be connected with the tourmaline in the Chickies quartzite.

Hornblende gneiss

General composition and origin of the hornblende gneiss.—The hornblende gneiss bodies throughout the schist are equally abundant in all metamorphic zones. An intrusive origin is suggested for the basic rocks since they occur as small lenticular masses, 20–50 feet thick, or in thin bands, some as narrow as 6 inches and all having sharp contacts with the schist. The dominant structure is gneissose, which is characteristic of the hornblende-rich rocks of the higher metamorphic grades of the amphibolite facies (Harker, 1933, p. 283–284). The association of hornblende with anorthite-bearing plagioclase in rocks of the amphibolite facies forms an assemblage which remains stable over a wide range of physical conditions (Turner, 1948, p. 76). This stability is indicated by the great similarity of composition in the hornblende gneiss throughout the various metamorphic zones (Table 10).

The main constituents are hornblende and andesine, plus a minor amount of quartz. In a few places some residual cores of pyroxene suggest a possible original gabbroic composition for the rocks.

In Table 11 a chemical analysis for specimen 47-P13a is given. The appearance and mode of this rock are typical of the hornblende gneiss in the area. The analysis is strikingly similar to that of Daly's (1933, p. 17) average plateau basalt.

Ungranitized hornblende gneiss.—The hornblende gneiss is a fine- to medium-grained mafic gneissose rock. All of the constituents show the effects of deformation by their arrangement in sub-parallel narrow bands. These bands are alternately felsic and mafic and are probably the result of mechanical differentiation during deformation and recrystallization of the rocks.

Accessory garnet occurs in only a few places, and the crystals are concentrated in bands parallel to the foliation. The conspicuous absence of garnet as an accessory in the hornblende gneiss may be due chiefly to the bulk composition of the original rock; the high metamorphic grades of the surrounding pelitic rocks indicate that

TABLE 10.—*Modes of ungranitized hornblende gneiss (arranged according to decrease in hornblende content)*

	Garnet zone			Staurolite zone	Kyanite zone			Sillimanite zone			Sillimanite zone								
	47-P13a	52-7*	47-P13h	52-284	52-61	52-236	52-250	47-P280a	52-368	52-13	52-350	47-P549	52-31	52-186	52-352	52-213	52-211	52-315	52-221
Pyroxene	—	—	26	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Hornblende	67	66	25	66	74	59	54	71	65	64	64	60	57	54	48	47	35	34	28
Plagioclase	30	31	33	31	21	14	17	20	18	10	29	20	37	22	29	29	38	46	24
Quartz	—	—	—	—	—	12	12	6	4	14	—	8	5	5	5	4	10	7	—
Biotite	—	<1	—	—	<1	—	17	—	3	—	—	9	—	14	16	14	14	6	8
Muscovite	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	10
Accessories	3	3	16	3	5	15	<1	3	2	12	7	3	3	5	2	6	3	7	30
Magnetite	X	X	X	X	X		X	X	X	X	X	X	X	X	X			X	
Sphene	X	X				X		X	X	X	X	X	X	X		X			X
Zircon	X		X				X						X		X	X			
Apatite				X		X	X	X		X				X		X	X	X	X
Epidote					X	X			X			X					X	X	X
Garnet										X								X	
Calcite											X								
Scapolite																X			
Sericite				X	X	X		X	X	X	X	X		X	X	X	X		X
Chlorite									X						X				X
	% An in Plagioclase																		
	31	26	36	29	38	45	49	40	45	42	40	40	40	35	36	42	50	34	38
	Grain size in mm.																		
Hornblende, length	0.05–0.5	0.005–2.0	0.5	0.02–0.05	0.02–0.5	0.05–0.3	0.05–0.1	0.05–0.2	0.01–1.5	0.02–0.4	0.2	0.1–0.5	0.05–0.2	0.1–0.3	0.1–1.0	0.3	0.05–0.3	0.01–0.2	0.2–0.5
Plagioclase, diameter	0.02–0.1	0.005–0.1	0.02–0.1	0.02–0.2	0.02–0.1	0.1	0.02–0.1	0.02–0.2	0.05–0.1	0.05–0.2	0.2	0.05–0.1	0.1–1.0	0.1–0.5	0.2–0.5	0.2	0.1	0.1–0.2	0.1–1.0
Quartz, diameter						0.05–0.3	0.05–0.2	0.005–0.05	0.01–0.5	0.02–1.0		0.05	0.05–0.1	0.05–0.3	0.02–0.4			0.05–0.3	
Biotite, length							0.05–0.2					0.1–0.5		0.05–0.3	0.2–0.5		0.02–0.1	0.01	0.2–0.7

TABLE 11.—*Chemical analyses of hornblende gneiss*

	47-P13a (Roberts Rd., Bryn Mawr College Power House)	47-T16 (½ mi. N. of Hulmeville	—* (½ mi. S. of Hulmeville)	—* (1 mile E. of Frankford)	Plateau Basalt (Daly, 1933)
SiO_2	48.68	52.36	62.53	74.24	48.80
TiO_2	1.69	tr.	1.00	0.36	2.19
Al_2O_3	14.37	16.37	12.04	13.71	13.98
Fe_2O_3	4.00	0.46	1.90	2.01	3.59
FeO	10.09	9.05	1.41	—	9.78
MnO	0.22	0.33	—	—	0.17
MgO	6.32	5.76	1.80	1.09	6.70
CaO	9.23	9.47	11.52	1.68	9.38
Na_2O	2.31	1.31	1.25	0.64	2.59
K_2O	0.47	3.20	2.57	4.84	0.69
H_2O	2.49	0.99	2.43	0.48	1.80
P_2O_5	0.29	0.73	0.20	0.26	0.33
FeS_2	—	—	2.13	—	—
	100.16	100.57	100.78	99.31	100.00
Norms					
Quartz	2.82	—	21.36	48.18	0.90
Orthoclase	1.16	18.90	15.57	28.36	3.89
Plagioclase	47.01 (An_{49})	42.69 (An_{53})	33.87 (An_{44})	10.24 (An_{48})	46.47 (An_{53})
Hypersthene	22.00	2.88	—	2.70	20.18
Diopside	12.45	—	9.72	—	16.57
Olivine	—	20.10	—	—	—
Sphene	—	—	—	0.98	—
Apatite	0.68	1.66	0.31	0.28	0.68
Corundum	—	—	—	5.00	—
Hematite	—	—	0.96	2.08	—
Magnetite	5.80	0.70	1.39	—	5.34
Ilmenite	3.19	—	1.98	—	4.26
Wollastonite	—	—	12.05	—	—
Modes					
Hornblende	67	33			
Plagioclase	30 (An_{51})	40 (An_{30})			
Microcline	—	6			
Biotite	—	6			
Muscovite	—	—			
Quartz	—	13			
Accessories	3	2			

*No specimens available, but locations of analyses obtained from Bascom's notes.
47-P13a. Analyst, W. F. Hillebrand (Bascom, 1905, p. 319).
45-T16. Analyst, F. A. Genth, Jr. (Florence Bascom's notes)
—*. Analyst, F. A. Genth, Jr. (Florence Bascom's notes)
—*. Analyst, F. A. Genth, Jr. (Florence Bascom's notes)
} unpublished analyses.

pressure-temperature conditions would permit the formation of garnet. Accessory epidote can be seen megascopically in some places.

Microstructures in the hornblende gneiss are gneissose and, in the felsic bands, somewhat cataclastic. In the less cataclastic felsic bands, the texture is granoblastic. In the mafic bands, hornblende grains are arranged in a more or less

parallel yet interlocking manner, creating a mosaic texture. In many rocks the hornblende, in the course of recrystallization, has included smaller grains of plagioclase and quartz, and is poecilobastic.

The pleochroic formula for the hornblende is the same throughout the area: X = yellow green, Y = green, and Z = blue green. Variation in the intensity of the colors from place to place may mean some slight change in the chemical composition of the hornblende. The hornblende shows no alteration to other minerals, even where it has been subjected to extreme crushing.

The plagioclase from hornblende gneiss in the garnet and staurolite zones is a calcic oligoclase or sodic andesine. In the kyanite and sillimanite zones, the plagioclase commonly is a more calcic andesine. Some faint reverse zoning of the plagioclase was noted in the sillimanite zone. However the change in composition of the plagioclase from zone to zone is not particularly marked.

Biotite is a common constituent of the hornblende gneiss from the sillimanite zone. It is more rare in the kyanite zone and is almost completely absent from the garnet and staurolite zones. The biotite is intergrown with the hornblende and both minerals appear to have crystallized together. The biotite is pleochroic with X = yellow or light green, and Y = Z = olive brown or olive green.

Inclusions of small euhedral to subhedral zircon, surrounded by pleochroic haloes, occur in both the biotite and the hornblende. Apatite is a usual but minor accessory. Magnetite occurs in anhedral grains, and in some places is accompanied by hematite. Subhedral sphene generally is concentrated in bands along the edges of hornblende grains. Colorless or pale green epidote is found in bands of euhedral crystals parallel to the general foliation of the rocks.

Hydrothermal alterations are very minor and local. In a few places there is change of some biotite to chlorite. Thin, cross-cutting veins of calcite, probably hydrothermally introduced, were observed in one hornblende gneiss body. In one thin section there was some alteration of plagioclase to scapolite.

Granitized hornblende gneiss (Table 12).—About a third of the hornblende gneiss bodies in the schist are granitized, and their distribution is limited to the sillimanite zone. The adjective granitized should be put in quotation marks, for the altered rocks are somewhat puzzling. Some granitized hornblende gneiss occurs near granodiorite (Pl. I, near Verreeville and Holmesburg), but other similarly altered hornblende gneiss may be found at some distance from rocks which could supply the material to affect the alterations. Moreover, the schist surrounding the granitized hornblende gneiss does not show the effects of granitization. For example, 45-T26, from schist three quarters of a mile southwest of Neshaminy Falls, is almost at the contact of one of the largest granitized hornblende bodies in the area. Yet the chemical analysis and mode (Table 2) of the schist suggest no addition of granitic material. Where did the material which altered the hornblende gneiss come from? Why did granitic additions, if such they were, confine themselves to the hornblende gneiss and not enter the surrounding schist? Field relationships and petrographic study thus far have failed to answer these questions. If granitization is any process by which rocks become granitic in appearance and composition, then the alteration of the hornblende gneiss to more felsic and granitic rocks may be called granitization.

The changes from ungranitized to granitized hornblende gneiss may be traced in the field within individual hornblende gneiss bodies. The felsic appearance of the granitized hornblende gneiss is due to a decrease in the proportion of hornblende, an increase in the per cent of quartz, and the addition of potash feldspar. The mafic constituents are aligned in discontinous, subparallel bands, which give the rocks a gneissose structure.

The microstructures of the granitized hornblende gneiss are faintly gneissose, or, with a greater proportion of felsic constituents, cataclastic.

Much of the plagioclase is zoned, with the rims of the grains more sodic than the cores. The average plagioclase becomes a little more sodic with increasing granitization, but the change is not very marked.

Some minor replacement by biotite was noted along the edges of the hornblende grains, but the hornblende and biotite are generally intergrown, as in the ungranitized hornblende gneiss.

TABLE 12.—*Modes of granitized hornblende gneiss*

TABLE 12.—*Modes of granitized hornblende gneiss*
(Arranged according to increase in microcline content)
(All in sillimanite zone)

	52-208	52-62	52-112	52-216	52-306	45-T16	52-217	52-102	52-219	52-218	52-220	52-176
Pyroxene	—	—	10	—	—	—	—	—	—	—	—	—
Hornblende	12	35	12	15	30	33	13	37	1	—	8	21
Plagioclase	60	46	51	50	36	40	44	33	21	31	29	17
Quartz	22	4	20	21	25	13	19	6	36	37	25	22
Biotite	1	6	<1	9	1	6	6	<1	13	1	3	2
Muscovite	—	—	—	—	—	—	—	—	—	1	—	—
Microcline	—	2	2	2	5	6	16	22	26	29	34	36
Accessories	5	7	5	3	3	2	2	2	3	1	1	2
Magnetite			X	X		X						
Sphene	X		X	X	X	X	X	X	X	X	X	X
Zircon	X	X	X		X			X			X	X
Apatite		X		X	X		X	X			X	X
Epidote	X	X			X	X		X	X		X	X
Allanite	X											
Zoisite			X							X		
Sericite	X	X		X	X	X	X	X	X	X		X
Myrmekite		X	X	X	X		X	X		X	X	X
Tourmaline								X				
Chlorite			X		X		X					

	% An in plagioclase											
	33	32	44	40	26	30	32	29	34	30	28	33
	Grain size in mm.											
Hornblende, length	0.05–0.4	0.02–0.2	0.05–0.2	0.7	0.1–0.5	0.7	0.05–0.2	0.5	0.5	0.5	0.05–0.2	0.1–0.5
Plagioclase, diameter	0.1–1.0	0.05–0.2	0.1	0.05–0.3	0.2	0.05–0.3	0.05–0.5	0.1–0.3	0.1–0.3	0.1–0.3	0.1–0.3	0.1
Quartz, diameter		0.02–0.1	0.1	0.05–0.1	0.01–0.1	0.05–0.1	0.05–0.1	0.05–0.2	0.05–0.2	0.05–0.1	0.05–1.0	0.02–0.05
Biotite, length	0.1–0.5	0.05–0.2	0.2	0.2–0.5		0.2–0.5	0.05–0.5	0.1–0.2	0.1–0.2	0.05–0.2	0.05–0.2	0.05–0.2
Microcline, diameter		0.05	0.1	0.1–0.2	0.1–0.2	0.1–0.2	0.1–0.2	0.05–0.2	0.05–0.2	0.1–0.2	0.1–0.5	0.1–0.5

With increasing granitization, the grains of hornblende and biotite become strongly anhedral. However, the occurrence of the biotite does not seem to be related to the granitization processes, since its proportion in the hornblende gneiss does not necessarily increase with increasing granitization. This fact is as puzzling as the lack of granitization in the surrounding schist.

Accessory epidote is commonly zoned. Alternating bands of isotropic or weakly birefringent allanite in epidote grains were noted at one place. The centers of some epidote grains are chloritized.

Myrmekite is present only in minor proportions. The plagioclase is more sericitized than in the ungranitized hornblende gneiss. Sphene is large and euhedral. Magnetite, zircon, and apatite are additional accessories.

Regional metamorphism of hornblende gneiss.—In regional metamorphism of basic rocks (Harker, 1933, p. 280–284) most of the mineralogic transformations, such as increase in the calcic molecule of the plagioclase, diminution of epidote with increase of sphene, and change of aluminous hornblende to green hornblende, have taken place by the time the metamorphism has reached the grade indicated by the garnet zone of the surrounding pelitic rocks. From this metamorphic grade upward, the main changes which can occur in the basic rocks are an increase in grain-size, a possible appearance of garnet, and, in the highest grade of metamorphism, a possible conversion of hornblende into pyroxene.

In this area, the hornblende gneiss, over a wide range of physical conditions, shows the stability characteristic of basic rocks of the amphibolite facies. There appears to be some slight increase in the calcic content of the plagioclase with increase in metamorphic grade. However, the hornblende gneiss does not become coarser from northwest to southeast, and there is no conversion of hornblende into pyroxene. Perhaps this indicates that the very highest intensities of regional metamorphism possible in the sillimanite zone had not yet been attained. The scarcity of accessory garnet, as has been suggested previously, is probably due to the initial composition of the basic rocks.

Granodiorites and Related Rocks

Granitization in the schist.—Along City Line from Darby Creek east to the Schuylkill River and along the Schuylkill River from City Line south to the Spring-garden Street Bridge 3 miles south of City Line, there are numerous occurrences of granitized schist. These rocks border the area of granodiorites and associated rocks described by Postel (1940). The appearance of the granitized schists is similar to that of rocks called by Postel (1940, p. 126) the intermediate replacement type of the Springfield porphyroblastic granodiorite. With an increase in the proportion of potash feldspar, occurring as lenticles or large porphyroblasts of microcline, the structure of the rocks, which may be called augen gneisses, grades from schistose to gneissose. No petrographic study was made of these rocks; detailed descriptions of similar rocks in the adjoining area are given in Postel's paper.

Granodiorites.—Whether or not the granodiorites in the region studied are related to the complex of rocks in the adjoining area to the southwest (Postel, 1940) will not be discussed here. Within the region limited by this paper, the granodiorites are small bodies, never more than a mile in length or half a mile in width, and are not abundant.

The granodiorites have a uniform appearance in the field. Some streaking or lineation, due to the orientation of the mafic constituents in subparallel discontinuous lines, creates a gneissose structure, indicating that the rocks were subjected to dynamic metamorphism after emplacement.

Cataclastic structure in these rocks can be seen in the field. The granodiorite at Chestnut Hill, less than a quarter of a mile south of the Rosemont fault, shows the results of marked cataclasis. The crushing may have been caused by an involvement of the rock with the movement along the fault, or may represent the movement producing the mylonite zones in the hybrid rocks just to the north of the schist (Armstrong, 1941). The contacts between the granodiorite and the schist are sharp wherever they are visible.

Thin sections show clearly the results of cataclasis. The milling down of the mineral

constituents is the most pronounced feature. There are excellent examples of flaser structure, especially in the granodiorite in Chestnut Hill. Many strain shadows in the quartz and plagioclase, together with the bent and distorted twin lamellae in the plagioclase and microcline, indicate crushing.

Myrmekitic intergrowths between the quartz and plagioclase grains are abundant. Accessories are subhedral to euhedral sphene, rounded or euhedral zircon, subhedral zoisite and clinozoisite, and subhedral magnetite.

The modes of specimens from all the granodiorites in the area are similar (Table 13). Very probably, these bodies of granodiorite are all from the same intruding magma.

Specimen 52-46, from the granodiorite in Chestnut Hill, has an accompanying chemical analysis (Table 14). The sharp contacts between the schist and the granodiorite, together with the uniform modal composition, indicate a magmatic origin.

Using Daly's (1933, p. 15) average granodiorite as a basis for comparison, the analysis of 52-46 reveals a higher percentage of silica and alumina and a lower percentage of lime and total iron. The norm of 52-46 has a higher proportion of quartz, an unusually low proportion of mafic constituents, and over 4 per cent of corundum. These facts could be accounted for by assimilation of some schist by the granodiorite during its emplacement.

TABLE 13.—*Modes of granodiorites*

	Vereeville			Holmesburg		Chestnut Hill		South of Byberry, Pennsylvania	
	52-124	52-120	52-121	52-145	47-C39	52-6	52-46	52-177	52-172
Quartz	36	28	30	35	46	48	49	35	24
Plagioclase	39	43	41	37	27	15	18	37	37
Microcline	15	17	18	20	22	23	20	21	33
Biotite	5	5	6	6	4	1	1	7	6
Muscovite	5	4	4	1	—	10	10	—	—
Accessories	<1	3	1	1	1	3	2	<1	<1
Magnetite	X		X		X			X	
Zircon	X	X	X	X	X			X	X
Sphene					X		X	X	X
Apatite	X	X	X				X		X
Garnet		X		X					
Clinozoisite							X		
Zoisite						X		X	
Chlorite		X						X	X
Myrmekite	X	X	X		X	X	X	X	X
Sericite	X		X	X		X	X	X	X
	% An in plagioclase								
	8-16	13	9	12-16	12-16	16	16	16	10-16
	Grain size in mm.								
Quartz	0.05–0.2	0.01–0.1	0.01–0.1	0.01–0.7	0.01–0.7	0.01–0.2	0.01–0.2	0.03–0.3	0.05–0.5
Plagioclase	0.1–0.5	0.1–0.3	0.1–0.3	0.1–0.3	0.1–0.3	0.01–0.3	0.01–0.3	0.2	0.05–0.5
Microcline	0.1–1.0	0.4	0.2	0.2–0.5	0.2–0.5	0.1–0.8	0.1–0.8	0.5	0.1–0.7

Pegmatites.—The pegmatites in the schist may be considered late-stage aqueo-igneous intrusions connected with the granodiorites in the area. The pegmatite bodies are from 5 to 100 feet thick and from 20 feet to almost a quarter of a mile long. Most pegmatites show some effects of crushing.

Cream-colored or flesh-pink microcline is the

TABLE 14.—*Chemical analyses of granodiorites*

	E. side of Wissahickon Creek, ¼ mile N. of Bell's Mill Road, Chestnut Hill 52-46*	World granodiorite† (Daly)
SiO_2	70.48	65.01
TiO_2	0.28	0.57
Al_2O_3	17.84	15.94
Fe_2O_3	0.99	1.74
FeO	0.68	2.65
MnO	0.03	0.07
MgO	0.28	1.91
CaO	1.89	4.42
Na_2O	3.58	3.70
K_2O	3.89	2.75
H_2O	0.52	1.04
P_2O_5	—	0.20
	100.46	100.00
Norms		
Quartz	30.54	20.10
Plagioclase	39.84 (An_{24})	49.79 (An_{37})
Orthoclase	22.79	16.68
Hypersthene	0.70	6.21
Diopside	—	2.26
Magnetite	1.16	2.55
Hematite	0.16	—
Ilmenite	0.61	1.06
Apatite	—	0.31
Corundum	4.59	—
	100.39	98.96
Mode		
Quartz	49	
Plagioclase	18 (An_{18})	
Microcline	20	
Muscovite	10	
Biotite	1	
Accessories	2	

* Phyllis Vipond, Analyst, Department of Chemistry, Bryn Mawr College

† Daly (1933, p. 15).

dominant feldspar, although oligoclase is present. Large crystals of microcline are common. Muscovite is in thick books. Quartz is grayish and generally transparent. Black prisms of tourmaline are conspicuous accessories.

Hydrothermally altered rocks.—Apart from the area of extensive granitization to the southwest, hydrothermal alterations are minor, but may be observed in all rock types.

Quartz veins, generally only 1–3 feet thick but in a few places as much as 20 feet thick, are the most obvious evidence of hydrothermal activity. The quartz veins are distinct from the numerous thin quartzose bands which are mechanical segregations produced during regional metamorphism. It is possible the quartz veins represent the late residues of the granodiorites and pegmatites which intruded the schist.

The occurrence of minerals of possible hydrothermal origin is limited to small areas along the fault zone and a few other places in the region. There are minor changes of biotite to chlorite, some sericitization of the plagioclase, and a few hydrous minerals described in the section on the mineralogy of the schist. However, the chemical analyses of the schist (Table 2) do not seem to reveal any large-scale introduction of hydrothermal material. It is possible that, aside from the intrusion of the quartz veins, the hydrothermal alterations noted were due to the local excesses of water released from the sediments during regional metamorphism.

Exposures at Morrisville, Penna., on the Delaware River, and near Emilie, Penna.

Rocks along the Delaware River, at Morrisville, Penna.—For almost a mile along the Delaware River, at Morrisville, Penna., there are exposures of rocks which differ in appearance from all others in this region limited. Bascom was undecided whether or not to include them in the Wissahickon schist (E. H. Watson, personal communication.) On the map of the areal geology of the Trenton Folio (Bascom, 1909b), the section along the Delaware River appears covered with gravels. However, the exposures are there, bordered on the north by Chickies quartzite, from which they may be separated by the continuation of the Cream Valley-Huntingdon Valley fault, and covered on the south by gravels. If the rocks are not part of the Wissahickon schist, they may be either granitic intrusions into the schist or, possibly, part of the Baltimore gneiss complex which appears just to the north of Chickies quartzite (Pl. 1).

The exposures are largely coarse-grained felsic granitic-looking rocks. Bascom, in her field notebooks and labels on specimens, called them granite-gneiss. In the central part of the exposures, for about 150 feet along the river, there

TABLE 15.—*Modes from exposures at Morrisville, Penna., on the Delaware River, and near Emilie, Penna.*

	At Morrisville, along the Delaware River				South of Emilie, Pa.			
	Granodiorite		Hornblende gneiss		Gabbro			
	52-68	45-T47	52-65	52-66	45-T20	45-T49a	45T49b	45-T52
Pyroxene	—	—	—	—	3	3	—	3
Hornblende	—	—	52	54	40	38	47	2
Tremolite	—	—	—	—	—	—	—	65
Quartz	39	31	<1	—	10	19	—	1
Plagioclase	27	40	40	31	46	20	34	20
Microcline	11	24	—	—	—	20	—	—
Biotite	6	—	2	10	—	—	6	—
Muscovite	2	—	—	—	—	—	—	—
Accessories	15	5	6	5	1	<1	13	9
Garnet		X						
Magnetite	X		X	X				
Pyrite	X							
Zircon	X				X			
Sphene					X	X		
Apatite					X	X	X	
Epidote	X		X	X			X	X
Clinozoisite								X
Chlorite	X	X					X	X
Myrmekite		X				X		
Sericite		X	X		X	X	X	
	% An in plagioclase							
	10	16	36	32	32	30	26-30	28-38
Grain size in mm. (diameter)								
Pyroxene					0.1	0.1–0.3		
Hornblende			0.1–0.5	0.2–0.5	0.1–0.2	0.05–0.3	0.2–0.8	0.05–0.1
Quartz	0.05–0.2	0.5–1.5			0.1–0.5	0.02–0.3		0.01–0.1
Plagioclase	0.05–0.1	0.2–0.5	0.1–0.5	0.2–0.5	0.1–0.3	0.1–0.3	0.2–0.6	0.05–0.1
Microcline	0.2–0.8	0.2–0.4				0.05–0.5		
Garnet		0.2						

are some thin micaceous bands, which give the rocks a gneissose appearance. No lineation was observed. Some micaceous bands are in tightly compressed and sheared folds. Seen in plan view, the folds appear to have been produced by an almost horizontal couple, and small drag folds indicate movement towards the west on the south side. The fold axes plunge steeply to the southwest. There are numerous large garnets, many of which have weathered out leaving the rocks with pitted surfaces.

The rest of the felsic exposures along the river are somewhat more massive. They consist of plagioclase feldspar, flesh-colored microcline, and lenticles or irregular and contorted stringers of grayish or bluish quartz. The intergrowths of quartz and feldspar give the rocks the appearance of a graphic granite. Blue quartz, characteristic of the Baltimore gneiss (Bascom, 1909b, p. 5), occurs also in lenticular masses several inches thick and 1–2 feet long.

At the northern end of the exposures there are several strongly gneissose basic masses from 10 to 60 feet thick. Here lineation plunges 22 degrees northeast. The basic masses resemble greatly the hornblende gneiss in the Wissahickon schist, except that they contain bands, both comformable and cross-cutting, of exceedingly coarse-grained feldspar and horn-

blende. The bands show crushing and shearing. Accessory epidote is abundant.

The felsic rocks have gneissose and cataclastic microstructures (Table 15). All of the constituents show the result of crushing. The modal composition of these rocks is that of a granodiorite. The plagioclase grains are chiefly remnants or patches in myrmekite. Epidote, like that in the hornblende gneiss, occurs as larger crystals concentrated in sub-parallel bands, or smaller inclusions in the plagioclase. There are some hydrothermal alterations of plagioclase to sericite and accessory garnet to chlorite.

In the mafic rocks, microstructures are cataclastic or faintly gneissose. Textures are granular or palimpsest decussate. The crushing of the hornblende gneiss, which lies in the projected fault zone, has not completely destroyed the original textures.

The pleochroism of the hornblende is strong. X = yellow green, Y = green, and Z = deep blue green. The plagioclase is sodic andesine, and contains many little stringers or vermicular inclusions of a slightly more sodic plagioclase. Epidote, occurring either as zoned euhedral crystals concentrated in bands parallel to the faint gneissosity of the rocks or as abundant small inclusions in the feldspar, may have two origins. The zoned crystals may be part of the original hornblende gneiss, and the small inclusions very possibly may be due to a reconstitution of the plagioclase, affected by the surrounding granitic rocks.

Gabbroic rocks near Emilie, Penna.—To the south and west of Emilie and Upper Emilie, Penna., gabbroic rocks occur throughout an area of almost 1½ miles long and half a mile wide. They were mapped by Bascom (1909b) as hornblende gneiss, but they are not like the rest of the hornblende gneiss bodies throughout the schist, and they bear a strong resemblance to the gabbroic rocks of the Baltimore gneiss complex just to the north of this region in Bucks County. Their occurrence to the southeast of the Wissahickon schist and along the general strike of the exposures at Morrisville seems to strengthen the possibility that they belong, along with the rocks at Morrisville, to the Baltimore gneiss complex. Both of these areas have been mapped with the Baltimore gneiss symbol, on which is superimposed a question mark, indicating doubt concerning the formation to which the rocks belong (Pl. 1). No contacts between the gabbroic rocks and the schist were observed and the structural relationships between the two rocks are not known. However, if these gabbroic rocks and the rocks along the Delaware River (described above) are part of the Baltimore gneiss complex, then they probably underlie the Wissahickon schist.

The rocks are in large rounded residual boulders, from 5 to 15 feet thick and 10 to 30 feet long. They occur through the swamps and woods on low ground, with gravels covering the high ground. Just northeast of Emilie, one structure reading was obtained; here the foliation strikes northeast-southwest. The rocks are massive, coarse-grained, and generally mafic. In a few masses a gradation into somewhat more felsic rocks was noted.

The microstructures are cataclastic or massive, and the texture is palimpsest ophitic. Though the mafic constituent is chiefly hornblende, small cores of diopside and augite have been noted in many hornblende grains.

The rocks appear, in many places, to have undergone granitization or hydrothermal alterations. In the more felsic rocks, microcline is an important constituent, and the plagioclase, a calcic oligoclase or sodic andesine, is zoned, with the cores of the grains more calcic and the rims more sodic. Some of the plagioclase shows considerable alteration to sericite. In one thin section, almost all of the hornblende is replaced by masses of tremolite. Chlorite partially replaces minor biotite. Sphene, apatite, and epidote are accessories.

Structure

Introduction

In the Wissahickon schist, the complete recrystallization of the original sediments, the intricate folding, and the absence of horizon markers either in the form of fossils or traceable lithological units present difficulties in interpreting the structure.

The amplitude of completely visible folds, in general, is small, and numerous readings on the smaller structures must serve as clues to the major regional pattern. The analysis of the regional structure is based on the most marked

features: schistosity, axial planes of folds, the movement indicated by drag folds, and the plunge of the fold axes and lineation. Definite changes in the structures were noted with change of metamorphic grade.

The most prevalent type of folding in the region is flexural-slip folding, which is common in metamorphosed rocks composed of dissimilar layers. In the relatively competent layers the movement of simple bending due to compression predominates, and the relatively incompetent layers yield by slipping. The amplitude of the folds is greater in the more competent layers than in the less competent layers (Knopf and Ingerson, 1938, p. 159–162). The micas follow the outline of the arch of the folds, and in many places the lower surface of the arch turns down toward the crest of the underlying fold.

That a previous shear folding, or slipping along transsecting cleavage planes, preceded the present type of folding is possible but not demonstrable. Any traces of such shear folding have been destroyed by subsequent or continuing deformation. Aside from the fault zone, the dominant schistosity does not cross-cut the lithology and the only approach to anything like axial-plane cleavage is in some isoclinally folded areas in zones of lower metamorphic grade. Throughout most of the area, the schistosity is parallel to the compositional layers and hence may be termed bedding schistosity. However, the S-plane of recrystallization cannot be considered S_1, or the plane of original crystallization parallel to the sedimentary beds. The S-plane is more probably the "compromise schistosity" suggested by Turner (1948, p. 280), which results from a rotation and gradual coincidence of the schistosity parallel to the compositional layering with the S_2 planes or transecting cleavage planes developed during earlier deformation. The compositional layers are nowhere continuous for long distances and, in many places, thin out and disappear; a slicing or shearing-process, followed by rotation and new folding, may have taken place. If the Wissahickon schist of this area could be traced to rocks of lower metamorphic grade than that of the garnet zone, or if the phyllite north of this area, in which the dominant schistosity is shear cleavage, proved to be a lower-grade equivalent of the Wissahickon schist, then the transition from shear or axial-plane cleavage to bedding schistosity could be seen. However, the equivalence of the Ordovician Octoraro phyllite and the Wissahickon schist has not been demonstrated, and the lowest metamorphic grade in the area studied is one of fairly advanced intensity of regional metamorphism; therefore, the previous existence of shear folding and axial-plane schistosity is an unproved possibility.

In regions of highest metamorphic intensity, rocks may begin to "flow in accord with their physical properties" (Cloos, 1937, p. 57–58). Such a process produces folds which are highly irregular and appear to have no definite directions. In flow folds the competence or incompetence of beds is obliterated. In the region studied, flow folds are not common, and have been observed locally only in the southern and southwestern portions of the sillimanite zone near intruding granodiorites. Probably the igneous rocks supplied the heat necessary to cause flowage.

Major Structural Features

Schistosity.—In general, the regional strike of the schistosity is northeast-southwest. The strike is very constant in the garnet, staurolite, and northeastern part of the kyanite zones. In the southwestern part of the kyanite zone, there are local deviations. Near intruding granodiorites in the southern portion of the sillimanite zone and near the migmatites to the southwest, there is a fairly wide divergence from the regional strike. However, throughout the rest of the sillimanite zone the schistosity strikes dominantly northeast-southwest.

In the garnet and staurolite zones, the schistosity, in general, is nearly vertical or dips steeply to the southeast. The dip is more variable in the kyanite zone, but in the sillimanite zone, particularly in the southern portions, schistosity dips more consistently to the northwest. In the sillimanite zone many dips are very steep, but just as many are quite gentle, and possibly these two groups of dips represent the two limbs of overturned folds.

Throughout most of the fault zone, the schistosity parallel to the compositional banding dips steeply to the southeast. (See readings near faults on Plate 4.) No actual fault plane was observed, but the dip of the schistosity in the fault zone possibly represents the dip of the fault itself.

The closely spaced cleavage planes, or dominant schistosity, in the garnet zone are crossed by a faint fracture cleavage, along which some recrystallization has occurred. This fracture cleavage may represent an involvement of the

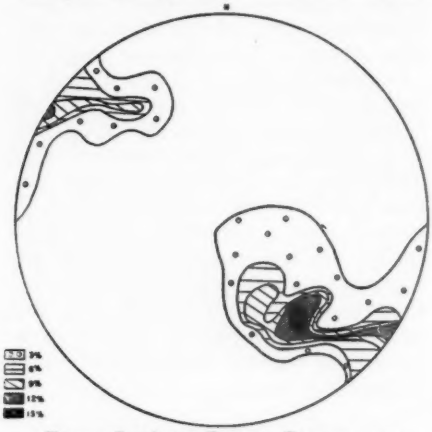

FIGURE 5.—AXIAL PLANES, GARNET AND STAUROLITE ZONES

33 poles, plotted on upper hemisphere. Overtuning of folds is dominantly to the northwest.

zone with the movement of faulting, since the garnet zone lies along and just to the south of the Rosemont fault. The fracture cleavage is more marked near the fault zone and dies out about a fourth to half a mile away from the fault zone.

Throughout the entire fault zone, the fracture cleavage is very noticeable and a fairly well developed schistosity occurs in a few places, cross-cutting the schistosity which parallels the compositional layers (Pl. 3, fig. 3). No definite fracture cleavage has been observed in the rest of the area, but in the staurolite zone the planes of schistosity are highly rippled or crinkled. Most of the ripples are small, from 1.0 mm. to 1.0 cm. across, but as the kyanite zone is approached, the size of the ripples increases and the schistosity surfaces resemble large logs piled one on the other (Pl. 3, fig. 5). The ripples may be caused by the breaking and rolling, during folding, of the more brittle felsic layers between the more plastic micaceous layers.

Folds.—The largest folds observable in the field are about 100 feet across. However, the existence of larger folds may be inferred from the changing dip in the schistosity from place to place. Most of the folds are considerably smaller. There generally is a strong similarity in the shape and attitude of all folds, large and small, in any one area.

FIGURE 6.—AXIAL PLANES, KYANITE ZONE

26 poles, plotted on upper hemisphere. Folds are overturned both to northwest and southeast.

No complete folds of any great size were observed in the garnet zone. The thin quartzose and micaceous bands have some minor plications or tight folds, only an inch or two in amplitude, which may represent the structural remnants of a previous deformation, but most of these folds have been destroyed by a slipping or shearing movement, probably connected with the faulting in the region. In the beginning of the staurolite zone, folding is tightly compressed and, in places, isoclinal. In the garnet and staurolite zones the overturning of the folds is predominantly to the northwest. The poles of 33 axial planes were plotted on an equal-area net (Fig. 5). The maximum shows axial planes dipping fairly steeply to the southeast. Similar diagrams were made for 26 axial planes in the kyanite zone (Fig. 6) and for 85 axial planes in the sillimanite zone (Fig. 7). In Figure 6, the maxima indicate fairly steep axial-planes, dipping southeast and northwest. The folding in the kyanite zone is more open than in the staurolite zone. Individual folded compositional layers may be traced for as much as 50–100 feet. Towards the southern part of the kyanite zone the folding becomes more generally over-

FIGURE 7.—AXIAL PLANES, SILLIMANITE ZONE
85 poles, plotted on upper hemisphere. Overturning of folds is dominantly to the southeast.

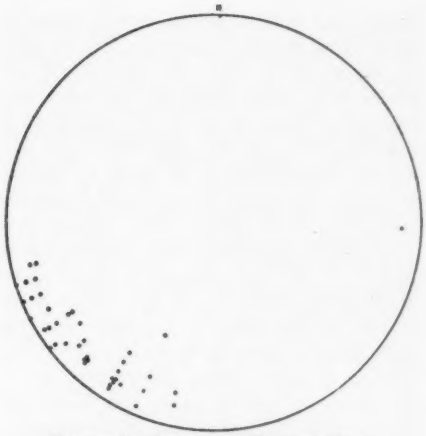

FIGURE 9.—LINEATION, KYANITE ZONE
42 points, plotted on upper hemisphere.

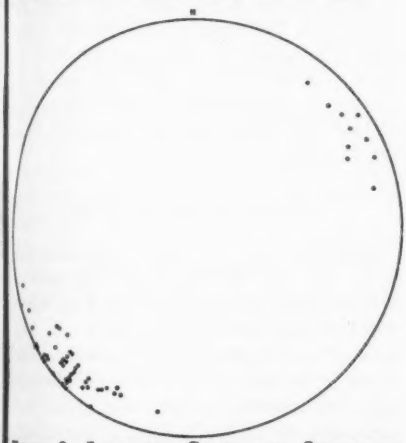

FIGURE 8.—LINEATION, GARNET AND STAUROLITE ZONES
58 points, plotted on upper hemisphere.

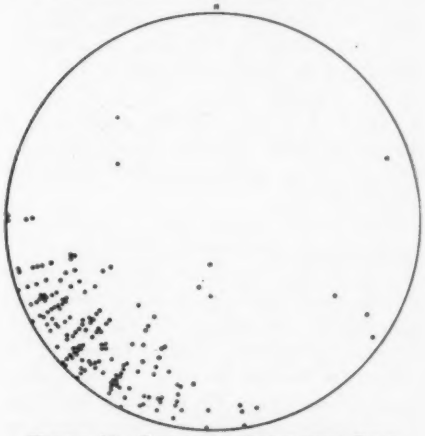

FIGURE 10.—LINEATION, SILLIMANITE ZONE
168 points, plotted on upper hemisphere

turned to the southeast. The folds in the sillimanite zone are dominantly overturned to the southeast (Fig. 7). Towards the southern part of the sillimanite zone, small recumbent folds, 1-8 feet in amplitude, seem to be impressed on an essentially vertical schistosity (Pl. 3, fig. 2) that may represent the steeper limbs of larger overturned folds.

A marked lineation, which generally represents the folds axes, is visible on the schistosity planes. The strike and plunge of the lineation and the fold axes for the garnet and staurolite zones, the kyanite zone, and the sillimanite zone are plotted (Figs. 8, 9, 10) on an equal area net. The average regional plunge in the garnet and staurolite zones is 8 degrees northeast. This gentle plunge is consistent with the strong parallelism of the strike of the schistosity in this part of the area. A minor concentration of points on the other side of the diagram represents a reversal of plunge to the southwest, noted in the fault zone from Glenside west to Chestnut Hill along the Cream Valley fault and along the Rosemont fault from Chestnut Hill southwest to Foxcroft. The reversal of

plunge may indicate several periods of movement during faulting (Armstrong, 1941, p. 685). In the staurolite zone, the average regional plunge is 11 degrees northeast. There is a corresponding small increase in the variation of the regional strike. In the sillimanite zone, the average regional plunge is about 16 degrees northeast. However, there is a much greater variability in the plunge and a fairly wide variability in the regional strike. In the southern part of the sillimanite zone, there is some steep-axis folding, particularly in the vicinity of intruding igneous rocks. It is possible that the steep-axis folding was produced by a differential reaction to stress, during deformation, of the more resistant igneous masses and the less resistant schist. In Figures 9 and 10, the minor reversals of direction of plunge represent lineation along the Cream Valley-Huntingdon Valley fault from Glenside east to Valley Falls. In some places along the fault zone, two lineations were noted on the schistosity planes. The more marked lineation plunges 40 to 45 degrees northeast. A fainter lineation, plunging 5–10 degrees northeast, cross-cuts the steeper lineation, and may represent repeated movement along the fault.

Drag folds or minor folds on the limbs of larger ones are more common in the micaceous than in the felsic layers. In general, the larger and the smaller folds are similar in outline and have parallel axial planes, though some minor local variations occur, probably because of the differential reaction to stress of the various compositional layers. The drag folds follow the rule of competent folding by showing movement up towards the anticlinal crests.

Along the Cream Valley-Huntingdon Valley and Rosemont faults numerous small drag folds show movement up to the north. Hence, the faults are believed to be high-angle thrusts, with the schist moving up relative to the northern-bordering Baltimore gneiss and Paleozoic sediments. The truncation of the zones of metamorphism by the faults (Pl. 1) indicates that the movement of faulting occurred either at the very end of or after regional metamorphism.

Minor Structural Features

Joints.—Not many joints were recorded. They occur chiefly in the less micaceous rocks, and the majority of those observed are normal to the fold axes. They are probably tension joints. In some places two cross-cutting joint systems make acute angles with the schistosity. A few longitudinal joints were noted. The greater number of joint planes are vertical or nearly vertical.

Boundinage.—Tension during folding produces boundinage or the breaking and separation of competent layers into fragments around which the incompetent material flows. Small-scale boundinage of the more brittle, felsic layers in the schist has been observed in many places (Pl. 3, fig. 4). The long dimension of the lenticular masses is parallel to the schistosity of the enclosing rocks.

The igneous intrusions, ultrabasic rocks, hornblende gneiss, pegmatites, and the granodiorites, are all in lenticular bodies oriented in more or less sub-parallel lines across the area (Pl. 1). This arrangement seems to suggest large-scale boudinage. It is possible that the igneous masses originally were in more or less continuous sheets which during folding were broken and drawn out into separate lenses (Pl. 3, fig. 6).

Structures in Intrusive Bodies

Ultrabasic rocks.—In the northwestern part of the area near the Rosemont fault there are five lenticular serpentine bodies from 100 to 500 feet thick and 500 feet to 1 mile long. Several small anthophyllite masses occur elsewhere in the area (Pl. 1). Since the ultrabasic rocks are conformable with the schist, and their structures are similar to those in the schist, it may be assumed that they intruded the schist before deformation. The origin and alteration of the ultrabasic rocks are subjects for a separate study.

Hornblende gneiss.—The hornblende gneiss bodies vary from small bands only several inches thick and a few feet long to large masses 200 to 500 feet thick and over a mile long. The larger masses are less common, however, and the average size of most hornblende gneiss bodies is 20–50 feet thick and 50–100 feet long. The very small masses do not show on the scale to which the map is drawn (Pl. 1).

The hornblende gneiss has been folded and deformed simultaneously with the deformation

of the schist, since the structures in the hornblende gneiss are conformable with the structures in the schist. Steeply dipping or vertical joints are common in the hornblende gneiss. The strike of most of the joints is perpendicular to the fold axes, though some joints have been observed to strike at acute angles to the foliation.

Granodiorite.—The centers of the granodiorite masses do not show intricate folding characteristic of the schist. However, discontinuous lenses of granodiorite, sheared off at the contacts of the schist and granodiorite, are folded conformably with the structures in the schist. Moreover, the gneissose structure of the granodiorite is parallel to the general schistosity of the schist, suggesting that both schist and granodiorite were subjected to a common deformation.

Jointing, similar to that in the hornblende gneiss, was noted in the granodiorite. At several places, horizontal faults show relative movements of a few inches.

Pegmatites.—The majority of pegmatites are conformable with the schist, and all contacts between the pegmatites and the schist are sharp. A xenolith of schist about half a foot thick and a foot long was observed near the border of a pegmatite. Some pegmatites show folding conformable with that in the schist.

In addition, cross-cutting pegmatites, 10–30 feet thick and 50–100 feet long are also folded and deformed. This suggests that there was a later deformation after the intrusion of the cross-cutting pegmatites, or that deformation, producing the major structures in the schist, continued through the emplacement of the cross-cutting pegmatites.

In the fault zone several intensely sheared pegmatites indicate that the faulting took place subsequent to their emplacement.

Interpretation of Regional Structure

In the Westchester and Coatesville quadrangles, 10–27 miles southwest of Bryn Mawr (Bascom and Stose, 1932, p. 4–6), the Wissahickon schist, as a part of the Glenarm Series, lies south of the Baltimore gneiss and stratigraphically above it. In Maryland (Knopf and Jonas, 1929, p. 152) the Baltimore gneiss is in domal uplifts, which are unconformably overlain by the Glenarm Series. The Baltimore gneiss can be traced from Avondale for about 60 miles east to Trenton, but in the Philadelphia area, the stratigraphic relationship of the Wissahickon schist to the Precambrian gneiss is obscured by a high-angle reverse fault which separates the two formations.

In the Philadelphia area, the general pattern of the rocks suggests a synclinal structure. The overturning of the folds to the northwest in the garnet and staurolite zones, the overturning of the folds to the southeast in the sillimanite zone, and the fact that the folds obey the laws of competent folding are compatible with the structures of an abnormal synclinorium (Leith, 1923, p. 165) produced by compression. Gruner (1941, p. 1630–1632) describes the formation of such a structure. A more competent crystalline floor is presupposed to underlie less competent sediments. During lateral compression, the more competent underlying rocks serve as buttresses between which the sediments are folded. The sediments dip steeply where they are in contact with the buttresses. In this area, the Baltimore gneiss flanks the Wissahickon schist on the north along the fault zone. If the exposures on the Delaware River at Morrisville, Pa., and to the south and west of Emilie, Pa., are part of the Baltimore gneiss, the more competent underlying rocks may have served as buttresses during compression and folding of the less competent overlying schist, and the structures in the schist could be just such as those described by Gruner.

An upbowing or sise in the central portion of a region, together with flow folding, could produce such an overturning of folds as has been noted in this area (Bain, 1931, p. 521). However, in such an anticlinal structure, the layers in the central portion would be either flat or very gently folded. In this area, the central portion of the structure is strongly folded. Moreover, the folding is flexural-slip folding and not flow folding. Hence, the attitude of the axial planes throughout the region is taken to represent a synclinal structure produced by compression, rather than an anticlinal structure in which there has been flow folding.

Along a traverse, from Chestnut Hill southeast for 4 miles along the east side of Wissahickon Creek (Fig. 11), almost continuous rock exposures reveal the pattern of regional folding.

The general trend of the major folds has been assumed from the appearance of the folding at the various outcrops and the attitude of the coincidence is merely fortuitous, or perhaps a larger regional study will yield similar correlations and possible explanation.

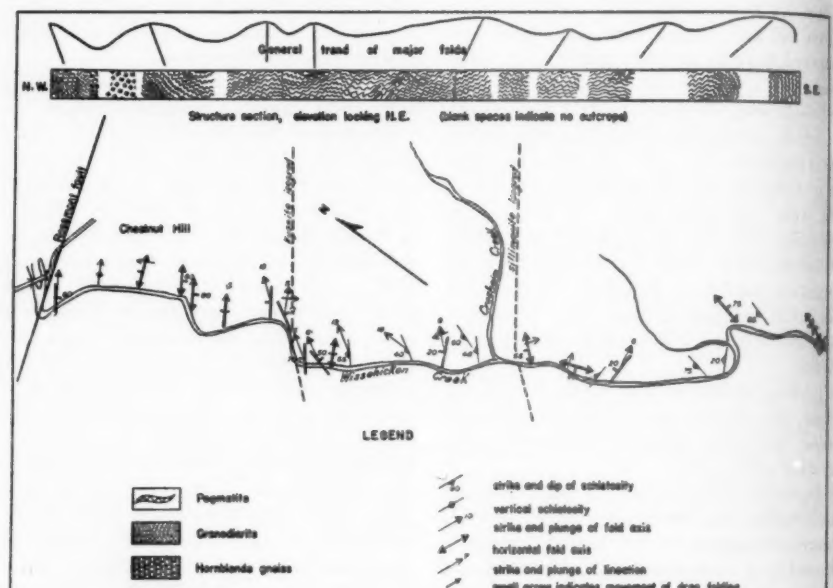

FIGURE 11.—TRAVERSE ALONG WISSAHICKON CREEK

axial planes. The beginning of the kyanite zone is roughly coincidental with the central portion of the major structure.

Structural control of metamorphism, though not readily explainable, has been observed in other regions (Billings, 1941, p. 923–924). In this area it is likewise difficult to explain, but nevertheless a fact, that the northern limb of the general synclinal structure coincides with the garnet and staurolite zone, the central portion with the kyanite zone, and the southern portion with the sillimanite zone. Perhaps the

SUMMARY AND CONCLUSIONS

Sandy, arkosic, and shaly sediments of varying thickness and composition were completely recrystallized during regional metamorphism which increased in intensity from northwest to southeast. Chemical analyses of sandstones, shales, and arkoses are similar to analyses of the schist. Hence it is inferred that, apart from regions to the southwest where there has been granitization, the metamorphism and recrystallization of the sediments into a schist occurred

without appreciable addition or subtraction of material. In this respect the metamorphism has been isochemical.

Increase in intensity of metamorphism caused the successive developement of garnet, staurolite, kyanite, and sillimanite, which occur in definite zones across the area. Within this region the rocks belong to the epidote-amphibolite and amphibolite facies.

The increase in the intensity of metamorphism was accompanied by changes in the character of the mineral constituents of the rocks. With increase in metamorphic grade the plagioclase became more calcic, the sodic oligoclase of the garnet and staurolite zones giving place to a medium oligoclase in the kyanite zone and a calcic oligoclase or sodic andesine in the sillimanite zone. Olive-green biotite, though occurring also in the zones of higher metamorphic grade, seems to be characteristic of lower metamorphism intensity. With increase in metamorphic grade orange-brown biotite becomes more predominant. The optical character of the garnet remains constant throughout all the metamorphic zones. The optical character of the staurolite remains constant in the staurolite, kyanite, and sillimanite zones. Sillimanite, in many places, appears to be pseudomorphous after kyanite. Accessory zoisite is more common in the zones of lower metamorphic grade. Accessory epidote is rare and confined, generally, to the sillimanite zone. Hydrothermally formed stilbite and chlorite occur locally as coatings on joint surfaces. Sericite is a minor and local alteration of plagioclase. Tourmaline is a late mineral formed after deformation.

There is a marked increase in grain-size with increase in the intensity of metamorphism. Texture is granoblastic and the microstructures are schistose or granulose in the zones of lower metamorphic grade. In the sillimanite zone, the microstructures are granulose or cataclastic. In this zone of highest metamorphic intensity, continuing deformation caused the breaking and partial crushing of the brittle felsic mineral aggregates.

The schistosity of the rocks is parallel to the compositional banding and hence is a bedding-schistosity, though the S-plane of recrystallization may be a compromise schistosity produced by the rotation and eventual coincidence, during deformation, of the S_2 or shear cleavage with the S_1 or plane parallel to the original bedding. S_2 planes or fracture cleavages were noted in the fault zones and in the garnet zone near the Rosemont fault. Fracture cleavage appears to die out within a distance of a fourth to a half mile from the faults.

The alteration of the hornblende gneiss to more felsic, granitic rocks is puzzling in view of the lack of corresponding granitization in the surrounding schist. This alteration of the hornblende gneiss has been called "granitization", for lack of a better term, and because, in its widest sense, granitization may embrace all processes by which rocks became granitic in appearance and composition.

Granitic gneissose rocks, different in appearance from all other rocks in the area, occur along the Delaware River, at Morrisville, Penna. Near Emilie, Penna., there are gabbroic rocks which strongly resemble gabbroic rocks in the Baltimore gneiss complex to the north of the Cream Valley-Huntingdon Valley fault, in Bucks County, Penna. These granitic and gabbroic rocks could be part of the Baltimore gneiss complex rather than of the Wissahickon schist. Since no contacts between these rocks and the Wissahickon schist were observed, the structural relationships between the two formations are not clear. If the rocks at Morrisville, Penna., and near Emilie, Penna., are part of the Baltimore gneiss complex, they probably underlie the Wissahickon schist.

From the pattern of overturning of folds and the competent character of the folding, the general regional structure is assumed to be an abnormal synclinorium developed by compression. It has not been possible to explain the coincidence of the garnet and staurolite zones with the northern limb of the synclinal structure, the kyanite zone with the center, and the sillimanite zone with the southern limb of the structure. Probably a large part of the northern limb of the synclinorium was cut off by the Rosemont and Cream Valley-Huntingdon Valley faults.

After the deposition of the sediments there was intrusion of basic igneous rocks. The structures and conformable relationship of most of the ultrabasic rocks with the schist suggest that they also intruded the schist before deformation. Granodiorites may have been emplaced after the basic rocks. The parallelism of the gneissose structure in the granodiorites and the schistosity

of the surrounding rocks indicates that they all were subjected to a common deformation. Pegmatites, considered to be late-stage aqueo-igneous intrusions connected with the introduction of the granitic rocks, are in both conformable and cross-cutting relationship with the schist. Folding and shearing of the cross-cutting pegmatites is evidence of continuing or later deformation. Faults to the north and northwest of the area truncate the major structure and the metamorphic zones of the schist. Hence, faulting represents the latest movement in the region.

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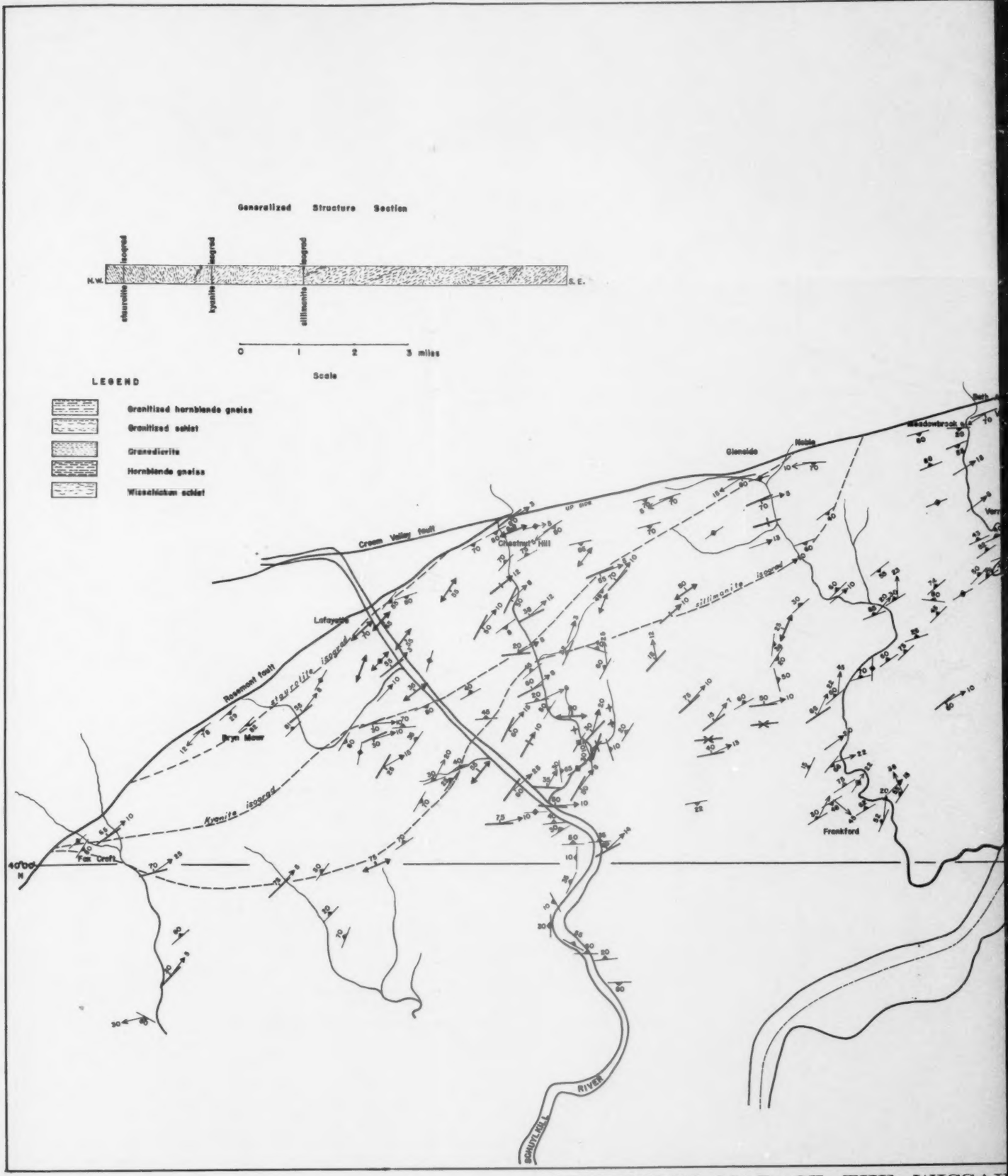
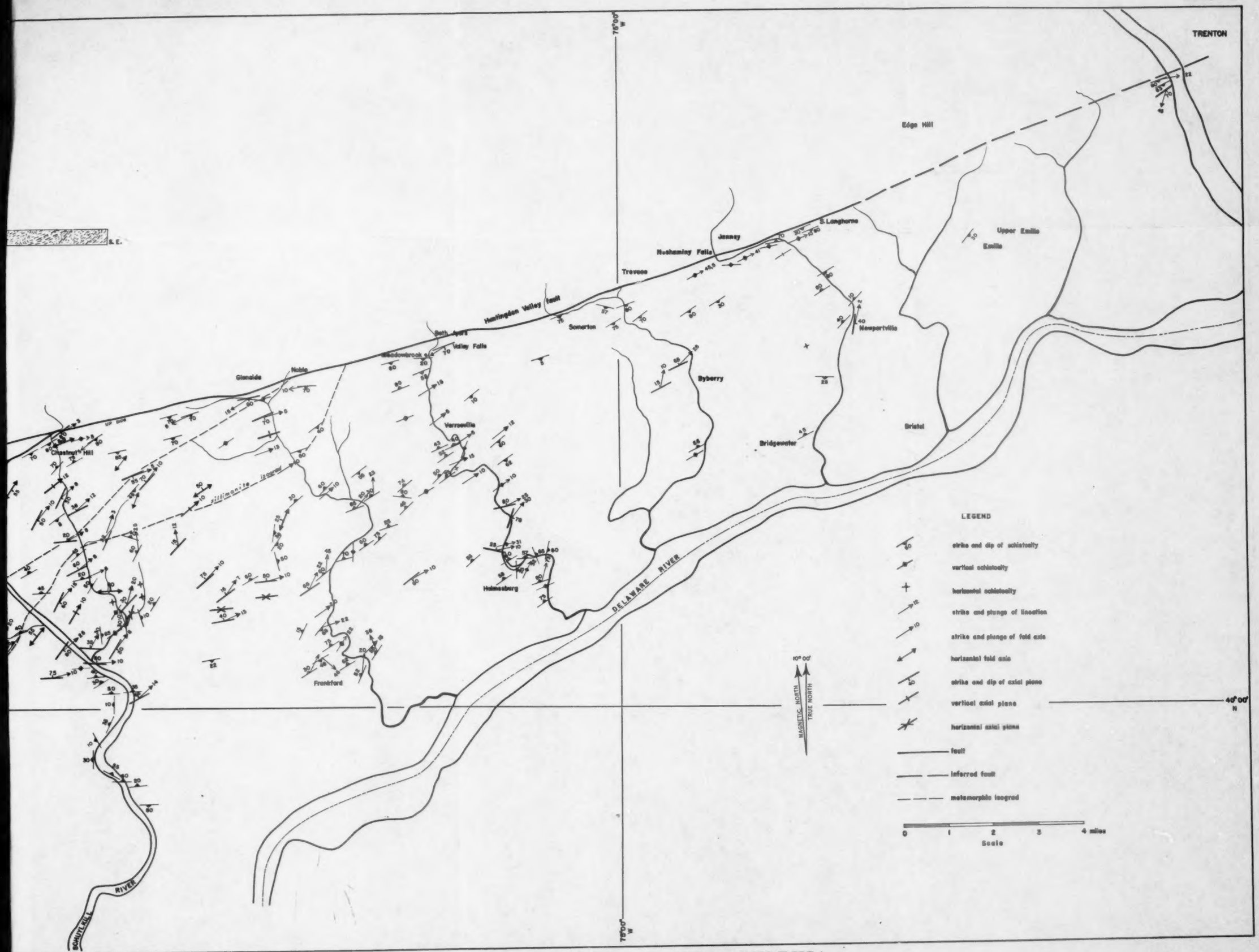


STRUCTURE MAP OF THE WISSAH



STRUCTURE MAP OF THE WISSAHICKON SCHIST AT PHILADELPHIA

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Short Notes

AGE DETERMINATIONS OF PEGMATITES IN TOPSHAM, MAINE

Two age determinations for pegmatites in Topsham, Maine, were discussed in a recent bulletin of the Maine Geological Survey (Shainin, 1948, p. 20). A specimen of samarskite from Topsham, presumably collected in a pegmatite, was analyzed by Gonyer (Lane and Marble, 1937, p. 59–60) by the lead-helium method, and gave approximately 232 × 10^6 years, or Carboniferous. Monazite from a pegmatite about a quarter of a mile from the samarskite locality was analyzed by Kroupa (Lane and Marble, 1937, p. 57), and the age of this specimen has been calculated by Adolf Knopf (Personal communications, November and December, 1947) as more than 600 × 10^6 years or Proterozoic. The present writer (1948, p. 20) stated that the Carboniferous age of Gonyer's analysis was more in harmony with the field relations at Topsham than the Proterozoic age of the other analysis.

It is of interest, therefore, to note in a subsequent paper by Ahrens (1949, p. 251) that his strontium-rubidium ratios indicate an age of 200 × 10^6 years for lepidolite from a pegmatite in Topsham. This tends to confirm the Carboniferous age reported by Gonyer.

Of course the possibility that more than one generation of pegmatite is present in Topsham must be considered. The present writer, however, could find no evidence to suggest the presence of more than one generation. On the contrary he was impressed with the similarity in internal structure, mineralogy, and degree of deformation in the pegmatites observed. These criteria, while they suggest a similar age for the pegmatites studied, do not, however, rule out the possibility that more than one generation of pegmatite is present.

Many age determinations are made of minerals from pegmatites. Investigators who publish these determinations would make their contributions much more valuable if they specified the rock type from which the mineral was derived and gave the precise location from which the sample was collected. In localities specified in the literature as "Topsham, Maine" and "Newry, Maine" (Ahrens, 1949), etc., more than a hundred pegmatites occur, and subsequent detailed studies may very likely prove the presence of more than one generation of pegmatite. Very little correlation between the geological history of an area and a reported age determination from that area can be made when the student finds the age of a mineral reported without indication of the rock type and exact locality.

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TWO PRESSURE VESSELS FOR SILICATE-WATER STUDIES

By O. F. Tuttle

The development of new alloys with greatly increased strength at elevated temperatures has opened a new field of research to the geologist. These alloys permit construction of simple pressure vessels that can be used to investigate silicate-water (or other volatile materials) equilibrium relations under pressure-temperature conditions considerably above those of the complete melting of granite. For example, at 15,000 psi water pressure Stone Mountain granite melts at approximately 750°C. whereas one of the pressure vessels to be described

has been used at 15,000 psi and *1050°C*. These vessels, then, serve to investigate equilibrium relations throughout the entire *p-t* range of

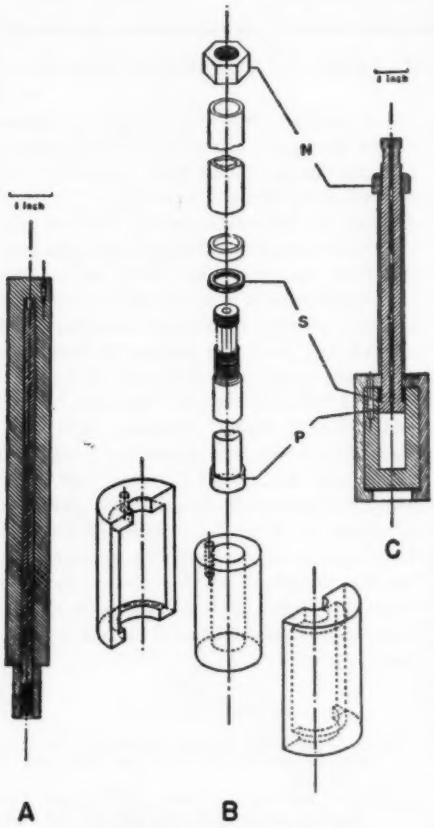

FIGURE 1.—PRESSURE VESSELS

(A) Cross section of cold-seal pressure vessel.
(B) and (C) Disassembled isometric drawing and cross section of pressure vessel utilizing the Bridgman seal. The isometric view shows the manner in which threads have been avoided by using a "split-cylinder" to maintain the seal.

the metamorphic rocks—from the greenschist facies through the sanidinite facies. The student of ore deposits can readily study problems of ore deposition under *p-t* conditions covering essentially all ore deposits from epithermal to pyrometasmatic. The mineralogist can utilize these pressure vessels to synthesize and study the stability range of minerals containing volatile components which heretofore could be investigated only by employing exceedingly complicated apparatus. Those silicates which have been difficult or impossible to crystallize from glasses in many cases can be readily crystallized in the presence of water at elevated pressures and temperatures. For example, albite glass can be heated dry for years without appreciable crystallization, whereas in the presence of 15,000 psi water pressure complete crystallization can be obtained to 10 to 20 minutes.

Research which previously required expensive and cumbersome equipment can be carried out with the aid of the new alloys much more readily and at considerably less expense. A set-up for research of this type at temperatures up to 1000°C. and pressures up to 30,000 psi costs less than a good petrographic microscope. One of the bombs (Fig. 1A) is simply a 1-inch rod of stellite 11 inches in length with a ¼ inch axial hole bored from one end to within ½ inch of the other end. The closed end is placed in a furnace to a depth of approximately 7 inches, leaving 4 inches outside the furnace. This end is only slightly above room temperature and can be connected directly to a pump which maintains the desired pressure, thus avoiding the problems encountered when the seal is in the hot part of the furnace. The charge is supported near the closed end which is adjusted in the furnace so that it lies in the position of maximum temperature. Convection and danger from explosive failure are minimized by having the support for the charge fill the greater portion of the internal volume of the bomb. A thermocouple is placed in a small well opposite the charge for the temperature measurements. This apparatus is called a "cold-seal pressure vessel" and has been used at 1050°C. with 15,000 psi water pressure. At somewhat lower temperatures pressures exceeding 45,000 psi (7 miles depth—density 2.7) can be utilized.

A second type of pressure vessel (Fig. 1B, C) of larger internal volume utilizes the Bridgman[1] seal and is used largely for the preparation of relatively large amounts of materials for subsequent study in the cold-seal bomb. It can also be used to run several samples of slightly different form or composition under

[1] Bridgman, P. W. (1931) *The physics of high pressure*, Macmillan, New York.

identical pressure-temperature conditions. The bomb closure utilizes the "unsupported area" principle, but the seal is maintained without the use of threaded parts, hence eliminating the thread seizure problem. Initially the seal is made by tightening the nut (*N*) which compresses the silver washers (*S*), expanding them against the wall of the bomb chamber. When water is admitted to the vessel by means of a pump the pressure against the plunger (*P*) is transmitted to the smaller area of the silver washers compressing them further and maintaining the seal. The seal is thus automatically tightened in proportion to the pressure maintained in the vessel.

The bomb has an internal volume of 16 cc. and has been used at temperatures up to 700°C. at 15,000 psi. As the design permits maintaining the seal at temperatures at which some permanent deformation occurs, considerably higher temperatures and pressures can probably be used.

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